A review of the NE Atlantic conjugate margins based on seismic refraction data

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Abstract: The NE Atlantic region evolved through several rift episodes, leading to break-up in the Eocene that was associated with voluminous magmatism along the conjugate margins of East Greenland and NW Europe. Existing seismic refraction data provide good constraints on the overall tectonic development of the margins, despite data gaps at the NE Greenland shear margin and the southern Jan Mayen microcontinent. The maximum thickness of the initial oceanic crust is 40 km at the Greenland–Iceland–Faroe Ridge, but decreases with increasing distance to the Iceland plume. High-velocity lower crust interpreted as magmatic underplating or sill intrusions is observed along most margins but disappears north of the East Greenland Ridge and the Lofoten margin, with the exception of the Vestbakken Volcanic Province at the SW Barents Sea margin. South of the narrow Lofoten margin, the European side is characterized by wide margins. The opposite trend is seen in Greenland, with a wide margin in the NE and narrow margins elsewhere. The thin crust beneath the basins is generally underlain by rocks with velocities of $>7\ \mathrm{km\ s^{-1}}$ interpreted as serpentinized mantle in the Porcupine and southern Rockall basins; while off Norway, alternative interpretations such as eclogite bodies and underplating are also discussed.

The opening of the NE Atlantic Ocean between East Greenland and NW Europe created mainly magma-rich margins, with exception of the northernmost area at the shear margin between Svalbard/SW Barents Sea and NE Greenland (Fig. 1). The magmatism is associated with the North Atlantic Igneous Province (Coffin & Eldholm 1994; Saunders *et al.* 1997), as evidenced by onshore flood basalts, basalt flows, seaward-dipping reflections (SDRs), volcanic centres and sills. The main phase of pre-break-up volcanism started at 61 Ma (Storey *et al.* 2007 and references therein). An increase in magma production rate at 56 Ma marks the transition to the syn-break-up volcanism (Storey *et al.* 2007).

Our knowledge of magma-poor rifted margins has significantly advanced over the last two decades, in particular as a result of the detailed studies along the Newfoundland–Iberia conjugate margin pair that involved the drilling of a number of scientific wells (see summary in Tucholke *et al.* 2007). In contrast, the understanding of magma-rich margins is still not satisfactory (cf. Quirk *et al.* 2014) as rift structures are overprinted by magmatism and seismic imaging problems are common. Seismic refraction data provide information on the large-scale crustal velocity structure. In particular, continental, transitional and oceanic crust can be distinguished, the amount of crustal thinning can be determined,

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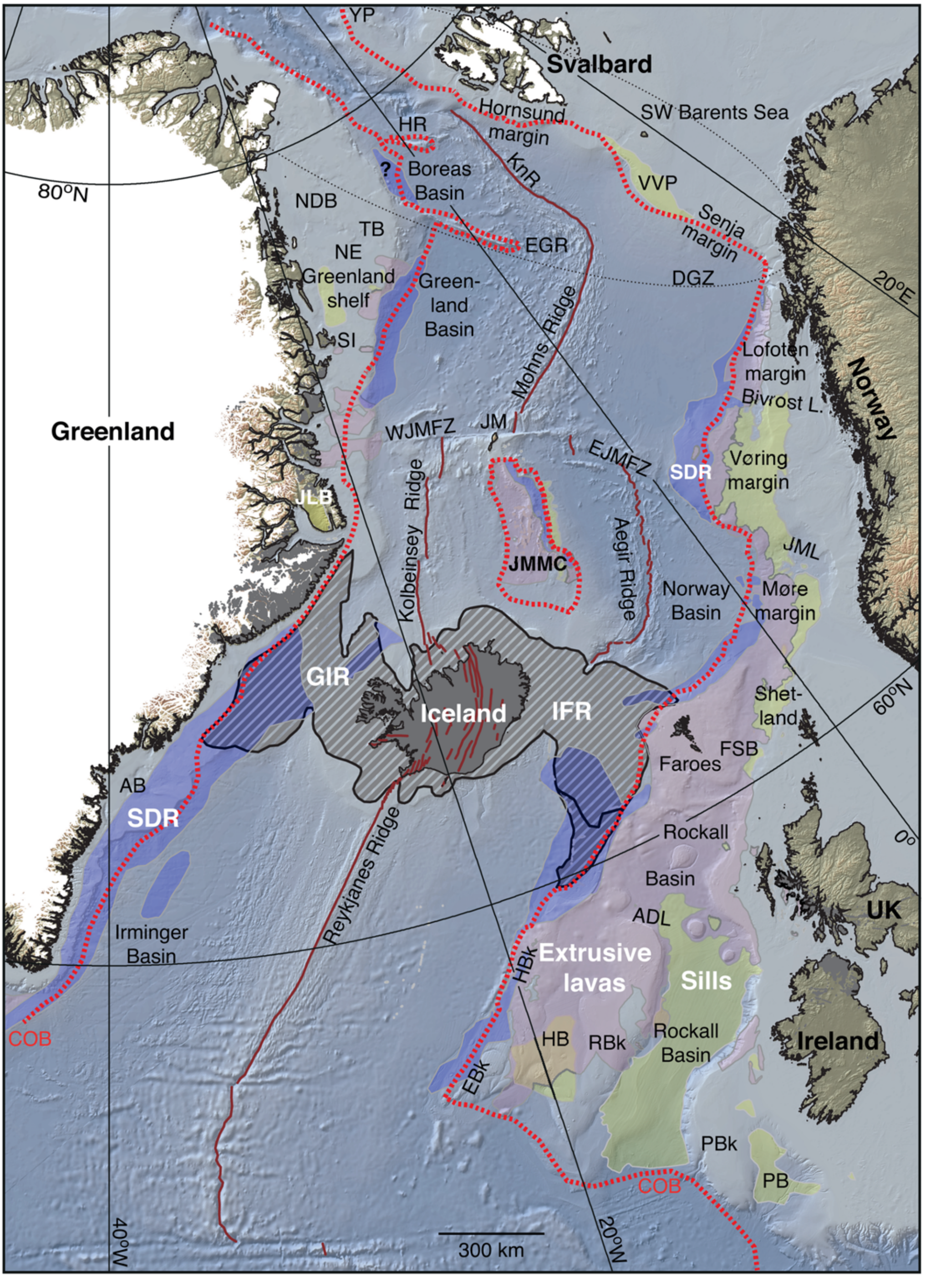


Fig. 1. Physiographical map (after Hopper *et al.*, this volume, in prep) of the NE Atlantic Ocean with the extent of the North Atlantic Igneous Province. The continent–ocean boundary (COB: dashed red line) is taken from Funck *et al.* (2014). Transparent overlays mark SDRs (blue), extrusive lavas (purple/pink), sills (green), the Greenland–Iceland–Faroe Ridge complex (hachured) and the onshore volcanic rocks (grey): all after á Horni *et al.* (2014), see also á Horni *et al.* (this volume, in review). Red lines mark the location of active and extinct spreading centres.

and magmatic additions can be recognized. However, when velocity models from conjugate margins in the NE Atlantic are compared, there are often inconsistencies. For example, the initial seafloor spreading between SE Greenland and the Hatton margin seems to be highly asymmetrical, with a higher spreading rate on the Greenland side (White & Smith 2009). If this asymmetry is real or just perceived depends largely on the correct identification of the continent–ocean boundary (COB). Funck *et al.* (2014) noticed that the various published COBs in the NE Atlantic deviate by as much as 150 km.

Despite a large number of published seismic refraction lines in the NE Atlantic realm, there is a lack of truly conjugate profiles (as derived from plate reconstructions such as the one in Gaina 2014 and Gaina *et al.*, this volume, in review). Often the lines are too short to image the entire margin from the proximal to the distal zone. When velocity models are stitched together, misfits at the intersection are very common and can result from differences in the modelling technique (e.g. forward modelling v. tomography), data density and quality, availability of coincident seismic reflection data, and the general concepts of margin evolution, which changed over time and vary between the working groups performing the modelling and interpretation. In addition, the experience of the modeller can also have some influence on the modelling results, as the proper identification of seismic phases is essential. Misfits at line intersections are easy to spot but become more of an issue when lines at conjugate margins are compared and conclusions on the margin development are drawn.

This paper reviews the seismic refraction data along the continental margins bordering the NE Atlantic Ocean and summarizes the current knowledge on the crustal structure. Conjugate margin transects are compiled to present the full rift system and to check for any inconsistencies in the models or interpretations. A full listing of the available seismic refraction lines in the study region is given in Funck *et al.* (2016, their fig. 2 and table 1). After a brief summary of the opening history of the NE Atlantic, this paper provides a more detailed account on the margin structure in three different subregions: the area south of Iceland; the Greenland–Iceland–Faroe Ridge; and the margins north of Iceland. In each section, the individual margin segments are described based on available seismic refraction lines, followed by a compilation of conjugate margin transects. The overall scope is to summarize the current knowledge of the crustal structure of the margins and to provide some ideas of where additional data would be beneficial to address some of the noticed inconsistencies in the reviewed datasets.

Geological setting

The NE Atlantic Ocean evolved from the Palaeogene break-up of an amalgamated landmass composed of Avalonia, Baltica and Laurentia that had formed the vast Caledonian mountain chain (Ziegler 1988). Prior to break-up, the landmass was subject to post-orogenic collapse in Devonian times, followed by several episodes of extension and rifting starting in the Carboniferous (Ziegler 1988; Doré *et al.* 1999). Magmatism associated with the North Atlantic Igneous Province began at 61 Ma and continued throughout the break-up phase (Storey *et al.* 2007), which resulted in magma-rich (volcanic-style) continental margins extending over a length of 2000 km on either side of the ocean (Fig. 1). Storey *et al.* (2007) estimated that continental break-up in the NE Atlantic occurred in the Eocene at 55.5 $\pm$0.3 Ma.

The margins are divided into a number of segments that correlate to a large degree with the segmentation of the present-day Mid-Atlantic Ridge (Fig. 1). South of Iceland, the SE Greenland continental margin is narrow and characterized by the absence of major rift basins, with exception of the poorly studied Ammassalik Basin (Hopper *et al.* 1998). On the conjugate Faroe–Hatton–Rockall margin, the Hatton and Rockall basins are the most prominent rift basins that together span a width of 600 km. The Hatton Basin terminates at the Anton Dohrn Lineament, which Kimbell *et al.* (2005) interpreted as a transfer zone. Across this zone, the axis of the Rockall Basin is shifted by 200 km.

Iceland is part of the Greenland–Iceland–Faroe Ridge (Fig. 1) where thick igneous crust in excess of 30 km is observed (Darbyshire *et al.* 1998; Richardson *et al.* 1998; Holbrook *et al.* 2001). The crustal thickness is the result of enhanced melting in the Iceland mantle plume (White & McKenzie 1995).

The opening history of the Norwegian–Greenland Sea between Iceland and the Jan Mayen

Fig. 1. (*Continued*) The dotted line indicates the region affected by the De Geer Zone (DGZ) megashear system. Abbreviations: AB, Ammassalik Basin; ADL, Anton Dohrn Lineament; EBk, Edoras Bank; EGR, East Greenland Ridge; EJMFZ, East Jan Mayen Fracture Zone; FSB, Faroe–Shetland Basin; GIR, Greenland–Iceland Ridge; HB, Hatton Basin; HBk, Hatton Bank; HR, Hovgaard Ridge; IFR, Iceland–Faroe Ridge; JLB, Jameson Land Basin; JM, Jan Mayen Island; JML, Jan Mayen Lineament; JMMC, Jan Mayen microcontinent; KnR, Knipovich Ridge; L., Lineament; NDB, North Danmarkshavn Basin; PB, Porcupine Basin; PBk, Porcupine Bank; RBk, Rockall Bank; SDR, seaward-dipping reflection; SI, Shannon Island; TB, Thetis Basin; VVP, Vestbakken Volcanic Province; WJMFZ, West Jan Mayen Fracture Zone; YP, Yermak Plateau.

Fracture Zone is characterized by numerous plate-boundary relocations that formed the highly or super-extended Jan Mayen microcontinent (JMMC) (Gaina *et al.* 2009; Gernigon *et al.* 2015). The initial opening occurred along the Aegir Ridge between the JMMC and the mid-Norwegian Møre Margin, and lasted until at least 30 Ma when the ridge became extinct (Jung & Vogt 1997; Gaina *et al.* 2009; Gernigon *et al.* 2015 and references therein). Gaina *et al.* (2009) related the fragmented character of the JMMC to several failed ridge-propagation attempts of the Kolbeinsey Ridge. By Chrons C6–C7 (25–20 Ma), stable seafloor spreading was established at the Kolbeinsey Ridge between East Greenland and the JMMC (Gaina *et al.* 2009).

Between the Jan Mayen Fracture Zone and the Greenland/Senja fracture zones, the East Greenland shelf widens northwards (Hamann *et al.* 2005), while the shelf off Norway with the Vøring and Lofoten margins narrows (Fig. 1). These two margins are separated by the Bivrost Lineament (Mjelde *et al.* 2003, 2005; Tsikalas *et al.* 2005).

Rifting in the NE Atlantic Ocean was linked to the Eurasia Basin in the Arctic by the De Geer Zone megashear system, encompassing the NE part of Greenland and the westernmost Barents Sea (Harland 1969; Mosar *et al.* 2002*b*). Off Norway, this region includes a transform zone off western Svalbard, and the sheared Hornsund and Senja margins (Fig. 1) (Eldholm *et al.* 1987; Engen *et al.* 2008). The strike-slip shear margin did not develop into a fully passive margin before the Oligocene (Faleide *et al.* 1996). The East Greenland Ridge is a sliver of continental crust that was sheared off the NE Greenland margin (Døssing *et al.* 2008; Døssing & Funck 2012). Another bathymetric high in this shear region is the Hovgaard Ridge, the crustal affinity of which is not yet fully resolved (Engen *et al.* 2008). North of the Greenland and Senja fracture zones, little volcanism is observed and restricted to the southernmost part (á Horni *et al.* 2014, this volume, in review). This includes the Vestbakken Volcanic Province on the Norwegian side of the margin, where sills intruded into Eocene sediments (Faleide *et al.* 1988; Ryseth *et al.* 2003). Conjugate to Vestbakken, some strong reflections observed in seismic data may indicate a volcanic cover (á Horni *et al.* 2014; cf. á Horni *et al.*, this volume, in review).

Continent–ocean boundary

Unequivocal oceanic and continental crust are separated by a continent–ocean transition (COT) zone, the composition and internal structure of which is often complex and ambiguous. At magma-poor margins, the nature of the basement in the transition zone is controversial, with models ranging from thinned and disrupted continental crust to exhumed mantle or ultra-slow spreading oceanic crust (cf. Peron-Pinvidic *et al.* 2013). At magma-rich margins, the interpretation of the COT becomes even more complicated as remnants of thinned continental crust may become indistinguishable from oceanic crust due to the break-up-related magmatic overprint. In the paper by White & Smith (2009), the COT was defined as the zone where a seaward increase in the average lower-crustal velocity is observed, indicating magmatic intrusions into older, seismically slower crust. With a dense spacing of seismic receivers, it is possible to determine the lower-crustal velocity distribution with sufficient resolution. However, many older seismic refraction lines do not have the necessary resolution to map the COT this way. At magma-rich margins, the initial oceanic crust generally has a higher lower-crustal velocity than normal oceanic crust, which White & Smith (2009) attributed to increased mantle temperatures; in the case of the NE Atlantic, the higher temperatures have been linked to the influence of the Iceland mantle plume.

It is often useful to define a distinct continent–ocean boundary (COB), outboard of which pure oceanic crust is present. This boundary is needed, for example, for plate kinematic reconstructions. In the following presentation of crustal velocity models along the NE Atlantic margins, the COB of Funck *et al.* (2014) is used (Fig. 1). As a starting point for the construction of the COB, Funck *et al.* (2014) mapped the landward limit of clear oceanic crust on seismic refraction lines. Initial refinements were then implemented on the basis of potential field data and other mapped structural features. Final adjustments were made after compiling plate reconstructions (Gaina 2014; cf. Gaina *et al.*, this volume, in review) for the time of break-up. In this paper, the COB is regarded as the seaward limit of the COT.

The plate reconstruction used in this paper is based on the identification of magnetic anomalies and fracture zones (Gaina 2014; cf. Gaina *et al.*, this volume, in review). In this reconstruction, the misfit of the COBs of conjugate plates at the time of closure is variable and displays some of the uncertainty in defining the COB. North of the Jan Mayen Fracture Zone up to the Senja margin (Fig. 1), the conjugate COBs of Funck *et al.* (2014) match within 20 km upon closure. The COBs of the northern JMMC and the Møre margin fit within 10 km, while the largest misfit of up to 70 km is observed south of Iceland.

Continental margins south of Iceland

This section starts with a review of the seismic refraction lines along the SE Greenland continental margin, followed by a description of the margins and rift basins in the Faroe–Hatton–Rockall region. Finally, three conjugate transects are presented and discussed with a focus on the observed asymmetries in crustal structure.

SE Greenland continental margin

The SE Greenland continental margin was studied in the SIGMA experiment, which was carried out in 1996 and consisted of four seismic refraction lines (Holbrook *et al.* 2001). They are all dip lines and three of them (SIGMA lines 2–4) are located south of the Greenland–Iceland Ridge (Fig. 2). The occurrence of thick igneous crust, including a high-velocity lower crust (HVLC: velocities well above 7.0 km s^{-1}) and a thick extrusive layer with systematically SDRs (Fig. 1), confirms that the SE Greenland margin is magma-rich along its entire length (Holbrook *et al.* 2001). The lateral changes along the margin are primarily related to the distance to the Iceland plume.

SIGMA line 2 (Korenaga *et al.* 2000) is located just to the south of the Greenland–Iceland Ridge and is characterized by 27 km-thick igneous crust at the location of break-up (Fig. 2a). The line is close to what Holbrook *et al.* (2001) defined as the proximal zone of the Iceland hotspot, which explains the increased crustal thickness. The oceanic crust remains relatively thick for another 50 km seaward before a noticeable thinning of the crust occurs. However, nowhere is the oceanic crust thinner than 8 km, which is more than the 7 km global average thickness of oceanic crust (White *et al.* 1992).

SIGMA line 3 (Fig. 2b) (Hopper *et al.* 2003) is at a more distal position to the Iceland plume and this is reflected by a further decrease in the oceanic crustal thickness at the break-up location (16 km). If the lower-crustal velocities are used as criteria for the definition of the COB (White & Smith 2009; see above), the COB may be located 60 km more landward than indicated by the COB of Funck *et al.* (2014). Irrespective of the location of the COB, the thinning of the oceanic crust is more gradual than along SIGMA line 2. Landward of the COB, the crust thickens and displays a maximum Moho depth of 33 km and a HVLC is observed.

SIGMA line 4 (Holbrook *et al.* 2001), at the southern tip of Greenland, is the most distal profile to the Iceland plume (Fig. 2c). The initial thickness of the oceanic crust at the break-up location (16 km) is not too dissimilar to what is observed on line 3. However, seaward, the crust thins to 8–9 km over only 70 km, which is narrower than what is observed on line 3 and indicates that the period of excess magmatism was shorter. Landward of the COB, the Moho deepens to 28 km and the crust displays a HVLC with a thickness of up to 8 km. Velocities of 6.7–7.3 km s^{-1} in the HVLC are not too well constrained, but the seismic records show prominent reflections from both the top and the base of the HVLC (J.R. Hopper pers. comm.). Extrusive volcanic rocks with velocities of >4.0 km s^{-1} can be correlated from the COB to the NW end of line 4, with the exception of a basement high south of Cape Farvel. Looking at the igneous crustal thickness along the SE Greenland margin, Holbrook *et al.* (2001) concluded that the initially wide melting anomaly (>1200 km) became restricted to the Greenland–Iceland Ridge in only 6 myr following the break-up.

SIGMA lines 2–4 did not detect any appreciable rift basins beneath the extrusive volcanics in the COT zone. The poorly defined Ammassalik Basin (Hopper *et al.* 1998; Gerlings *et al.*, this volume, in review) seems to be restricted to the area between SIGMA lines 2 and 3 (Funck *et al.* 2016). Line 3 is located along legs 152 (Saunders *et al.* 1998) and 163 (Larsen *et al.* 1999) of the Ocean Drilling Program (ODP). Here some meta-sedimentary rocks of unknown age were drilled at ODP site 917 (Larsen *et al.* 1998).

Faroe–Hatton–Rockall region

Similar to the SE Greenland continental margin, the Faroe–Hatton–Rockall region (Fig. 1) is characterized by high levels of pre-, syn- and post-break-up magmatism, as evidenced by SDRs, lava flows and sills (Hopper *et al.* 2014; á Horni *et al.*, this volume, in review). The Hatton, Rockall and Porcupine basins form the main rift basins in this area. Along RAPIDS line 4 in the Porcupine Basin (Fig. 3d), an up to 12 km-thick sedimentary sequence is underlain by continental crust that is extremely thin, in some places less than 2 km (O'Reilly *et al.* 2006). The three sedimentary layers are interpreted as Cretaceous and Cenozoic strata on top of predominantly Jurassic syn-rift deposits (O'Reilly *et al.* 2006). Mantle velocities beneath the basin centre are as low as 7.2 km s^{-1} and indicate a partial serpentinization of the mantle rock (O'Reilly *et al.* 2006). These velocities would be also compatible with magmatic underplating but O'Reilly *et al.* (2006) argued against this alternative for several reasons, such as the lack of a double reflection from the top and base of an underplated body. Beneath the adjacent Porcupine Bank and the Irish Shelf, the Moho deepens to 30 km, which is similar to full-thickness continental crust beneath Ireland and the UK (Funck *et al.* 2016). The crustal thinning to either side of the basin is asymmetrical, with a narrower necking zone in the west. O'Reilly *et al.* (2006) interpreted this to be the result of simple shear along low-angle westwards-dipping detachment surfaces formed during later stages of extension.

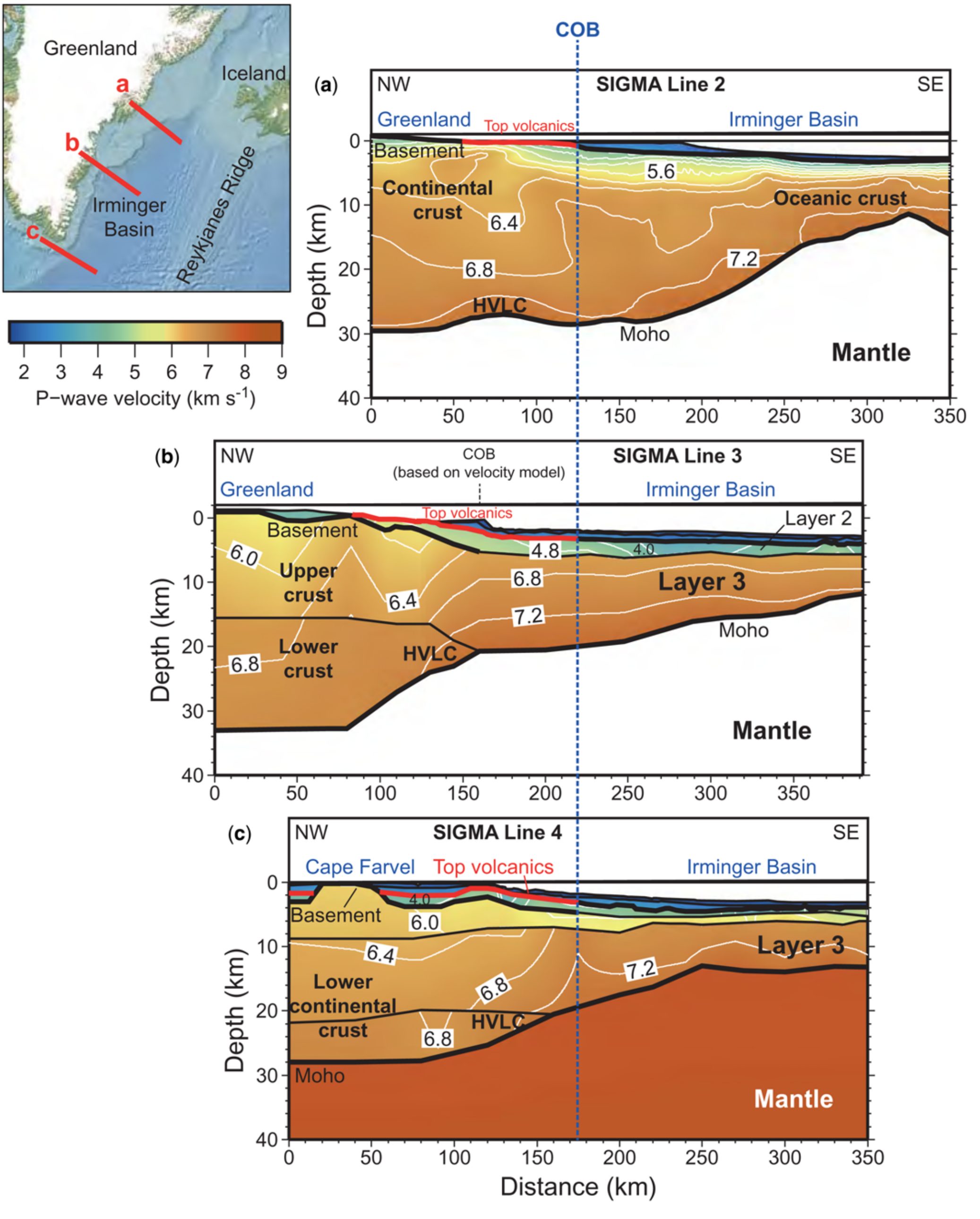


Fig. 2. P-wave velocity models at the SE Greenland margin: (**a**) SIGMA line 2 (after Korenaga *et al.* 2000); (**b**) SIGMA line 3 (after Hopper *et al.* 2003); and (**c**) SIGMA line 4 (after Holbrook *et al.* 2001). Velocities are given in km s^{-1} and the contour interval (white lines) is 0.4 km s^{-1}. Black lines mark layer boundaries in the velocity models. Abbreviations: COB, continent–ocean boundary; HVLC, high-velocity lower crust.

Serpentinized mantle is also interpreted beneath the southern Rockall Basin: for example, along RAPIDS line 1 (Fig. 3c) (Shannon *et al.* 1999). However, mantle velocities do not decrease below 7.5 km s^{-1}, indicating a lower degree of serpentinization (Christensen 2004) when compared to the Porcupine Basin. This is consistent with a thicker continental crust beneath the Rockall Basin. Along

RAPIDS line 1, the continental crust is stretched to a thickness of 5–7 km in a 200 km-wide zone (Shannon *et al.* 1999).

The amount of crustal thinning in the Rockall Basin decreases northwards. On AMP line E (Fig. 3a) (Klingelhöfer *et al.* 2005) in the northern Rockall Basin, the minimum crustal thickness is 12 km, twice as much as in the south. No mantle refractions are observed on AMP line E and there is no indication of mantle serpentinization on the cross-line (AMP line D) (Klingelhöfer *et al.* 2005). The sedimentary cover sequence on AMP line E is about 5 km thick, including up to 1.5 km of Palaeogene volcanic rocks. The sedimentary succession in the southern Rockall Basin (Fig. 3c) is typically 4.5 km thick and reaches a maximum of 7 km (Mackenzie *et al.* 2002).

To the west, the southern Rockall Basin is bounded by the Rockall Bank (Fig. 3c) that is characterized by a three-layered continental crust with a total thickness of 28–30 km (Bunch 1979; Shannon *et al.* 1999). The Hatton and Rockall banks are separated by the Hatton Basin. Available refraction data (Fig. 3c) indicate that the basin is underlain by continental crust with a thickness of about 15 km (Vogt *et al.* 1998; White & Smith 2009). The basin contains 1–3 km of post-break-up sedimentary rocks, but the deeper basin-fill is poorly understood because of the masking effects of the Cenozoic lavas. However, it appears likely that these lavas are underlain by Mesozoic rifts that contain rocks with relatively high velocities, because of either overcompaction or a high concentration of sills. Magnetic anomalies over the Hatton Basin suggest substantial volumes of magmatic rocks (Kimbell *et al.* 2010).

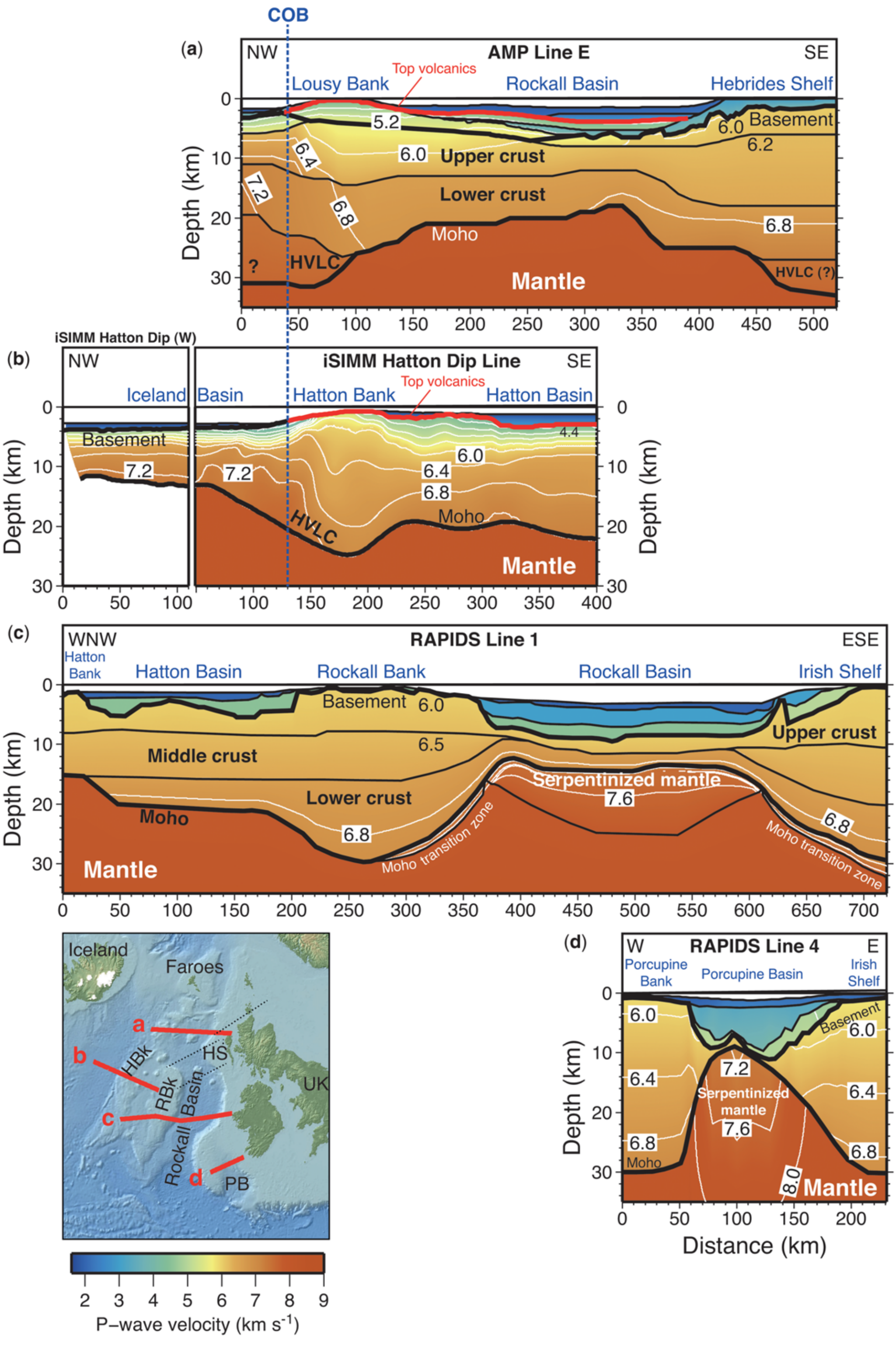
The outer continental margin extends from the Faroe Islands to Edoras Bank, linking a series of bathymetric highs (Fig. 1). The thickness of the crystalline crust beneath these highs is generally between 20 and 25 km (Funck *et al.* 2008; White & Smith 2009), but is slightly less (17 km) beneath Edoras Bank (Barton & White 1997). All these bathymetric highs are capped by several kilometres of volcanic rocks (Funck *et al.* 2008) (see also Fig. 3a). In places, low-velocity zones were detected between the volcanic strata and the underlying basement (Funck *et al.* 2008; Eccles *et al.* 2009), which may indicate the presence of sedimentary rocks predating the volcanism. The volcanic strata and SDRs are the result of voluminous pre-, syn- and post-break-up magmatism. Intrusive rocks and a HVLC are observed in a 40–50 km-wide transition zone (e.g. AMP line E and iSIMM Hatton dip line in Fig. 3a, b). The initial thickness of the oceanic crust is high, but decreases from north to south. At Lousy Bank, AMP line E (Fig. 3a) indicates that initial oceanic crustal thickness may be as great as 28 km (Klingelhöfer *et al.* 2005), further south at Hatton Bank it is 17 km (iSIMM Hatton dip line in Fig. 3b) (White & Smith 2009) and at Edoras Bank it is about 14 km (Fig. 4c) (Barton & White 1997).

North of Hatton Bank, the published seismic refraction lines are not long enough to fully map the decrease in oceanic crustal thickness over time. The best dataset available for the margin is the iSIMM Hatton line (Fig. 3b), where the oceanic crust thins from 17 to 9 km over a distance of 70 km to remain at that thickness to the end of the line (Parkin & White 2008; White & Smith 2009).

Conjugate transects south of Iceland

For all three dip lines (SIGMA lines 2–4) covering the SE Greenland continental margin, seismic refraction profiles could be found at the Faroe–Hatton–Edoras margin that are conjugate within 100 km (Fig. 4). The southernmost transect (Fig. 4c) is composed of SIGMA line 4 (Holbrook *et al.* 2001) and line CAM77 (Barton & White 1997). The two lines are joined at their position at magnetic Chron C22N (49 Ma), where both basement and Moho depth match each other. Up to a distance of 120 km away from Chron C22N, the Moho depth is fairly symmetrical. However, when the COB of Funck *et al.* (2014) is used, the zone of oceanic crust is roughly 20 km wider off Greenland. The initial oceanic crustal thickness on SIGMA line 4 is 16 km, while it is 14 km on line CAM77. While this deviation can be explained within the modelling uncertainties, the symmetry in Moho depth may, instead, argue for an incorrect position of the COB. Gaina *et al.* (2009) used a COB that is 35 km more landward at the Edoras Bank compared to the COB of Funck *et al.* (2014) used here. This indicates that the uncertainty in the position of the COB may well remove the apparent asymmetry in crustal accretion along this transect. However, it should be noted that basalts drilled at DSDP site 552 show evidence of contamination by either ancient U-depleted continental crust or subcontinental lithosphere (Merriman *et al.* 1988). This well is located about 10 km seaward of the COB proposed by Gaina *et al.* (2009).

Line CAM77 is too short to image the full width of the margin including the Hatton and Rockall basins. Therefore, RAPIDS line 1 (Vogt *et al.* 1998; Shannon *et al.* 1999) is shown in Figure 4c as a landward extension of line CAM77. The two lines are separated by 140 km, which may explain the difference in the crustal thickness where the models are joined. However, there is no isostatic balance between the Hatton Bank and the Hatton Basin for the model of RAPIDS line 1, which may


COB
(a)
NW
AMP Line E
SE
Lousy Bank
Top volcanics
Rockall Basin
Hebrides Shelf
Basement
Upper crust
Lower crust
Moho
HVLC
?
Mantle
HVLC (?)
Depth (km)
(b)
iSIMM Hatton Dip (W)
NW
iSIMM Hatton Dip Line
SE
Iceland Basin
Hatton Bank
Top volcanics
Hatton Basin
Basement
HVLC
Moho
Mantle
Depth (km)
(c)
WNW
RAPIDS Line 1
ESE
Hatton Bank
Hatton Basin
Rockall Bank
Rockall Basin
Irish Shelf
Basement
Middle crust
Upper crust
Serpentinized mantle
Lower crust
Moho
Moho transition zone
Mantle
Depth (km)
Iceland
Faroes
HBk
HS
RBk
Rockall Basin
UK
PB
(d)
W
RAPIDS Line 4
E
Porcupine Bank
Porcupine Basin
Irish Shelf
Basement
Serpentinized mantle
Moho
Mantle
Depth (km)
Distance (km)
P–wave velocity (km s-1)

indicate potential issues with the velocity model. The full transect shows three rift axes located within the Rockall Basin, the Hatton Basin and at the later break-up position. The Rockall Basin is fairly symmetrical, with similar necking profiles beneath the adjacent Rockall Bank and the Irish Shelf. Crustal stretching factors of 5–6 have thinned the crust sufficiently to allow for a partial serpentinization of the underlying mantle rock (O'Reilly *et al.* 1996). In the narrower Hatton Basin, the stretching factor is around 2. Final break-up occurred further to the west and is associated with a massive volcanic overprint of the crustal structure.

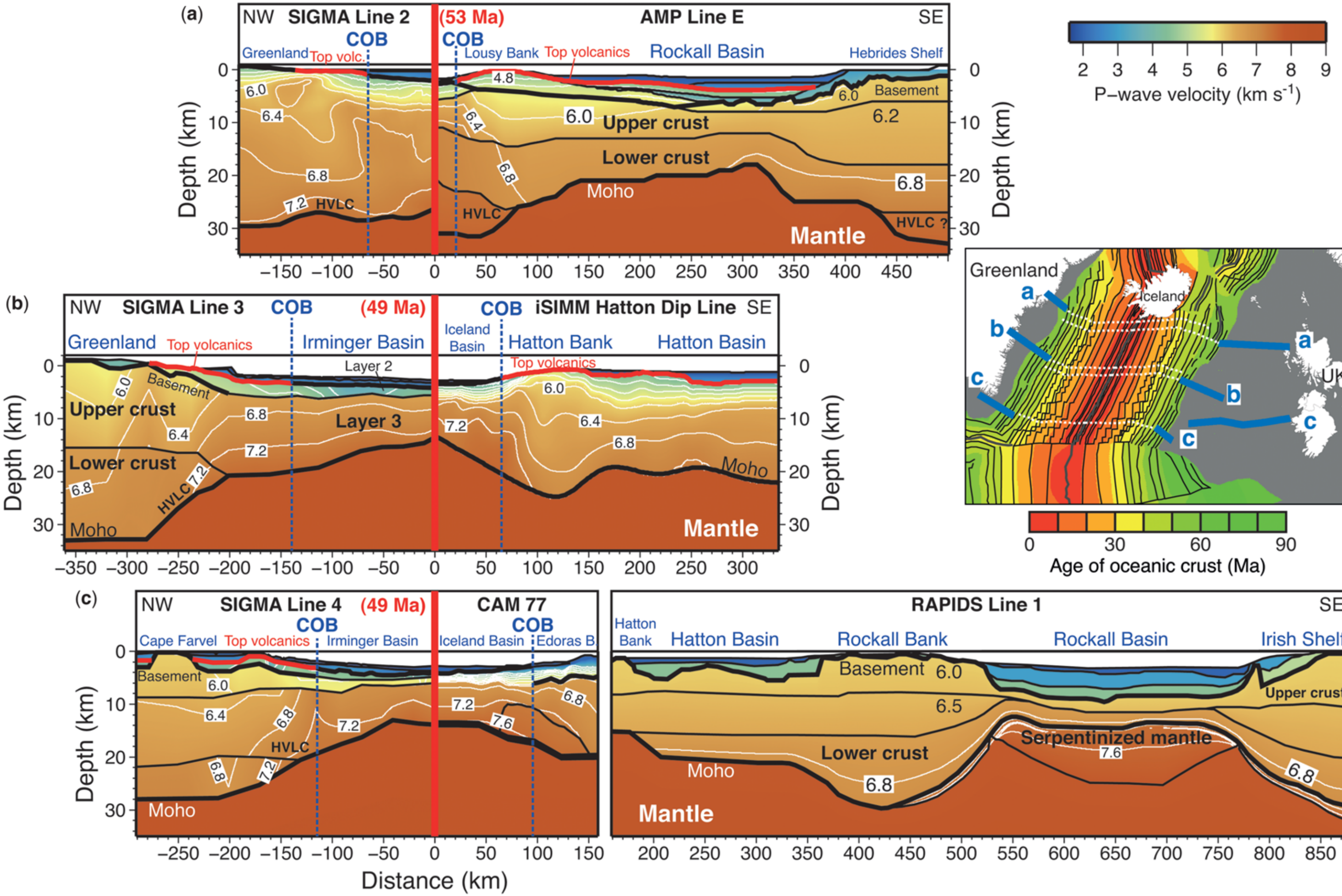
SIGMA line 2 (Korenaga *et al.* 2000) and AMP line E (Klingelhöfer *et al.* 2005) form a conjugate transect just to the south of the Greenland–Iceland–Faroe Ridge (Fig. 4a). The two lines are conjugate within 100 km and are stitched together at magnetic Chron C24N (53 Ma). AMP line E was primarily designed to determine the crustal velocity structure beneath the northern Rockall Basin, which is why the data coverage in the oceanic domain is limited. The width of the zone with inferred oceanic crust is, again, wider on the Greenland side than on the conjugate NW European margin. Given the deviation in the published COBs at either margin (cf. Funck *et al.* 2014), it is difficult to judge whether this asymmetry is real or not. In addition, a potential ridge jump in this region prior to Chron C21 (Kimbell *et al.* 2005) may further complicate this issue.

The initial thickness of the oceanic crust matches well, with 27 and 29 km on SIGMA line 2 and AMP line E, respectively (Fig. 4a). Volcanic rocks and HVLC are interpreted in the COT of either line. The transect covers the entire width of the conjugate margin pair, extending from onshore Greenland to full-thickness continental crust beneath the Hebrides Shelf. The crustal velocity structure between the two lines is difficult to compare owing to the different modelling approaches. The model of SIGMA line 2 is based on tomography, in which both the crust and any overlying volcanic rocks are modelled as one continuous layer (Korenaga *et al.* 2000). In contrast, the model for AMP line E is divided into several layers (Klingelhöfer *et al.* 2005), which distinguishes both a volcanic cover and up to four layers in the crystalline crust. However, both lines show a HVLC in the COT. Beneath the Hebrides Shelf, another 5 km-thick HVLC is indicated in the model based on an interpreted double reflection from the top and base of this layer. However, other lines in this region (their location is indicated by dotted lines on the map in Fig. 3) do not corroborate the presence of a HVLC (Powell & Sinha 1987; Roberts *et al.* 1988; Morgan *et al.* 2000). Volcanic rocks can be correlated from the top of Lousy Bank through to the east side of the Rockall Basin, where they become unrecognizable. However, some isolated lavas are known on the Hebrides Shelf (á Horni *et al.* 2014, this volume, in review).

The last conjugate transect along this margin segment is composed of SIGMA line 3 (Hopper *et al.* 2003) and the iSIMM Hatton dip line (White & Smith 2009) (Fig. 4b). The two lines are conjugate within 60 km and are joined at Chron C22N (49 Ma). Similar to the previous transect, the two velocity models were developed by different techniques. On the Hatton line, the volcanics and the crust were modelled as one continuous layer in the tomographical model, while the crust at the Greenland margin is divided into sublayers constrained by forward and inverse modelling. Despite these differences, the crustal thickness and velocity range match fairly well where the lines join. The crustal thinning on the European side of the transect is distributed over a much wider zone than on the Greenland side, as already discussed for the previous transects. The thickness of the continental crust beneath Hatton Bank (<24 km) is less than the full-thickness crust of Greenland (32 km). While these asymmetries are not surprising given that the margin was subject to several rift episodes preceding the final break-up (Ziegler 1988; Doré *et al.* 1999), the model shows a rather pronounced asymmetry in the accretion of oceanic crust.

The zone between Chron C22 and the COB interpretation of Funck *et al.* (2014) is 65 km wide at the Hatton margin, while it is 140 km wide at the SE Greenland margin. When the original interpretation of the SIGMA line 3 velocity model is used, the COB on the Greenland side may be even 60 km further landward (Hopper *et al.* 2003), which would increase the asymmetry even more. To avoid some of the problems associated with such a gross asymmetry at a spreading centre, White & Smith (2009) looked into possible alternative interpretations of the COB. In particular, they noted that some rather weak magnetic anomalies off Greenland that are

Fig. 3. P-wave velocity models of selected seismic refraction lines in the Faroe–Hatton–Rockall region: (**a**) AMP line E (after Klingelhöfer *et al.* 2005); (**b**) iSIMM Hatton dip line (W) (after Parkin & White 2008) and iSIMM Hatton dip line (after White & Smith 2009); (**c**) RAPIDS line 1 (after Vogt *et al.* 1998; Shannon *et al.* 1999); and (**d**) RAPIDS line 4 (after O'Reilly *et al.* 2006). Dotted lines indicate the location of the seismic refraction profiles on the Hebrides shelf (HS) with no evidence of the presence of a HVLC (Powell & Sinha 1987; Roberts *et al.* 1988; Morgan *et al.* 2000). Abbreviations: COB, continent-ocean boundary; HBk, Hatton Bank; HS, Hebrides Shelf; HVLC, high-velocity lower crust; PB, Porcupine Basin; RBk, Rockall Bank.



(a)
NW
SIGMA Line 2
(53 Ma)
AMP Line E
SE
COB
COB
Greenland
Top volc.
Lousy Bank
Top volcanics
Rockall Basin
Hebrides Shelf
Basement
Upper crust
Lower crust
Moho
Mantle
HVLC
HVLC ?
Depth (km)
P–wave velocity (km s-1)
(b)
NW
SIGMA Line 3
COB
(49 Ma)
COB
iSIMM Hatton Dip Line
SE
Greenland
Top volcanics
Irminger Basin
Iceland Basin
Hatton Bank
Hatton Basin
Layer 2
Basement
Upper crust
Layer 3
Lower crust
HVLC
Moho
Mantle
(c)
NW
SIGMA Line 4
(49 Ma)
CAM 77
COB
COB
Cape Farvel
Top volcanics
Irminger Basin
Iceland Basin
Edoras B
Basement
HVLC
Moho
RAPIDS Line 1
SE
Hatton Bank
Hatton Basin
Rockall Bank
Rockall Basin
Irish Shelf
Basement
Upper crust
Serpentinized mantle
Lower crust
Moho
Mantle
Distance (km)
Greenland
Iceland
UK
Age of oceanic crust (Ma)

interpreted as magnetic cryptochrons 24.1n–24.11n (Larsen *et al.* 1994; Larsen & Saunders 1998) might not be true seafloor spreading anomalies. Instead, White & Smith (2009) proposed that these anomalies could represent the edges of subhorizontal lava flows. If this model is true, the COB would be located further seaward than indicated by Funck *et al.* (2014). A remodelling of the SIGMA line 3 dataset may provide some more definitive answer as to where the COB is located if a similar modelling approach is used, as for the iSIMM Hatton dip line. The Monte Carlo tomography used on the Hatton line is better suited to map the lateral velocity variation across the COT than the forward modelling applied to the SIGMA line.

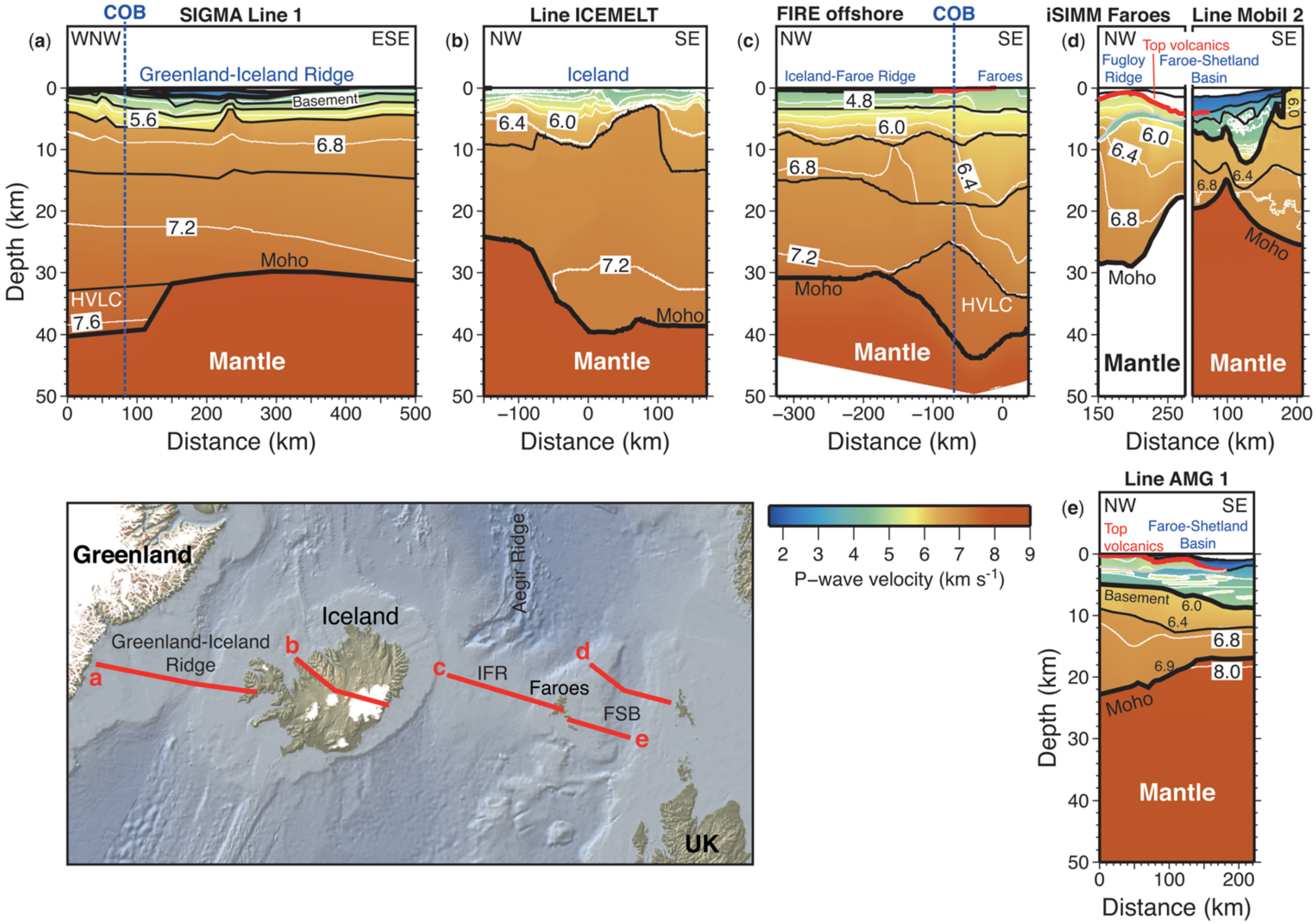
Some support for the reinterpretation of the weak magnetic anomalies on SIGMA line 3 by White & Smith (2009) may come from the conceptual break-up model at volcanic margins proposed by Quirk *et al.* (2014). This model is based on deep seismic reflection data from the NE Greenland margin and appears to provide a mechanism for generating asymmetry during the early stages of break-up. In this model, lava-filled half-graben develop on either side of an axial horst and there is no reason why these have to be symmetrical. In addition to the asymmetry, the model provides a means of generating broad zones of subaerial SDRs above crust that looks more oceanic than continental. Eventually, the rising asthenosphere causes the horst to split and spreading is accommodated by a sheeted dyke system. At this stage, an outer high is predicted, and there is a change to submarine conditions and a more symmetrical spreading. Applied to SIGMA line 3 (Fig. 2b), there is no prominent outer high. However, a zone with rough basement separates subaerial SDRs from submarine SDRs according to the interpretation of Hopper *et al.* (2003). The seaward limit of the subaerial SDRs occurs some 50 km seaward of the COB proposed by Funck *et al.* (2014). Moving the COB by this amount would greatly reduce the asymmetry in the initial seafloor spreading (Fig. 4b).

Greenland–Iceland–Faroe Ridge

The Greenland–Iceland–Faroe Ridge (GIFR) is composed of thick igneous crust and extends from the eastern coast of Greenland to the Faroe Islands (Fig. 5). The 25–35 km-thick crust of the GIFR is the product of the interaction between the Mid-Atlantic Ridge and the Iceland mantle plume (White 1997). White (1997) showed that an increase in mantle temperature of as little as 50°C causes an increase of 30% in crustal thickness. The three main tectonic elements of the GIFR are the Greenland–Iceland Ridge, Iceland and the Iceland–Faroe Ridge. The two seismic refraction lines on the crest of the Greenland–Iceland Ridge and the Iceland–Faroe Ridge (Fig. 5a, c) show a fairly constant Moho depth of around 30 km (Richardson *et al.* 1998; Holbrook *et al.* 2001), but there is some indication in the data that the Moho deepens close to the shore, although this is only poorly constrained on the Greenland side. The velocity model of SIGMA line 1 (Fig. 5a) indicates a fairly homogeneous velocity structure along the entire length of the mapped Greenland–Iceland Ridge with no clear hint as to where the COB could be located. The COB of Funck *et al.* (2014) is located at km 82. Other authors place the COB close to the coast of Greenland (e.g. Escher & Pulvertaft 1995; Mosar *et al.* 2002*a*), which would be more compatible with the velocity model of SIGMA line 1.

Similarly, the FIRE offshore line (Fig. 5c) provides only limited resolution of the lower-crustal velocity structure close to the Faroe Islands. Richardson *et al.* (1998) presented two models, one with and one without a high-velocity lower-crustal layer. Both models fit the seismic observations, and the deepening of the Moho towards the Faroes is robust. However, the maximum Moho depth is 4 km shallower in the model without the high-velocity layer. Richardson *et al.* (1998) preferred the model with the high-velocity lower crust (this is the model shown in Fig. 5c) based mainly on gravity modelling and on the amplitude characteristics of the Moho reflection. This model would also be consistent with nearby lines that show a high-velocity lower crust (e.g. the iSIMM Faroes line: Roberts *et al.* 2009). Crustal velocities gradually decrease towards the Faroes, starting some 100 km away from the islands. This argues for the presence of continental crust beneath the Faroes (Richardson *et al.* 1998). However, the resolution of the data is not sufficient to determine the COB based on the lower-crustal velocity distribution. Published

Fig. 4. P-wave velocity models along conjugate transects south of Iceland: (**a**) SIGMA line 2 (after Korenaga *et al.* 2000) and AMP line E (after Klingelhöfer *et al.* 2005); (**b**) SIGMA line 3 (after Hopper *et al.* 2003) and iSIMM Hatton dip line (after White & Smith 2009); and (**c**) SIGMA line 4 (after Holbrook *et al.* 2001), line CAM77 (after Barton & White 1997) and RAPIDS line 1 (after Vogt *et al.* 1998; Shannon *et al.* 1999). The inset map displays the age of the oceanic crust (after Gaina 2014; cf. Gaina *et al.*, this volume, in review), together with the line locations (blue lines) and flow lines (dashed white lines). Thin solid lines mark selected isochrons. Abbreviations: B., Bank; COB, continent–ocean boundary; HVLC, high-velocity lower crust; volc., volcanics.


SIGMA Line 1
COB
(a)
WNW
ESE
Greenland-Iceland Ridge
Basement
5.6
6.8
7.2
Moho
HVLC
7.6
Mantle
Depth (km)
Distance (km)
Line ICEMELT
(b)
NW
SE
Iceland
6.4
6.0
7.2
Moho
Mantle
Distance (km)
FIRE offshore
COB
(c)
NW
SE
Iceland-Faroe Ridge
Faroes
4.8
6.0
6.8
6.4
7.2
Moho
HVLC
Mantle
Distance (km)
iSIMM Faroes
Line Mobil 2
(d)
NW
SE
Top volcanics
Fugloy Ridge
Faroe-Shetland Basin
6.0
6.4
6.8
Moho
Mantle
Mantle
Distance (km)
Greenland
Greenland-Iceland Ridge
Iceland
Aegir Ridge
IFR
Faroes
FSB
UK
P–wave velocity (km s⁻¹)
Line AMG 1
(e)
NW
SE
Top volcanics
Faroe-Shetland Basin
Basement
6.0
6.4
6.8
6.9
8.0
Moho
Mantle
Depth (km)
Distance (km)

COBs in this area vary by as much as 50 km (cf. Funck *et al.* 2014).

Beneath Iceland, the Moho depth varies between 15 and 40 km (Funck *et al.* 2016). The maximum value is found on the ICEMELT line (Fig. 5b) (Darbyshire *et al.* 1998) in central SE Iceland above the postulated centre of the Iceland plume (Darbyshire *et al.* 2000). However, there is an alternative explanation for the presence of thick crust beneath that region. Foulger & Anderson (2005) proposed the presence of a microplate that may contain oceanic crust submerged beneath younger lavas. Recently, Torsvik *et al.* (2015) suggested the presence of a sliver of continental crust beneath SE Iceland that links with the JMMC. This interpretation is based on geochemistry data. The postulated continental fragment is covered by the eastern end of the ICEMELT line (Fig. 5b) (Darbyshire *et al.* 1998) and the onshore portion of the FIRE profile (Staples *et al.* 1997). Neither of these lines shows a distinct lateral velocity anomaly that could point to the presence of continental crust. Given that the continental crust may have been thinned significantly, followed by substantial magmatic addition, its velocity structure might be indistinguishable from Iceland-type igneous crust. This often-used term comprises the abnormally thick crust formed above the centre of the Iceland mantle plume, the surface of which was originally subaerial (White 1997).

The break-up-related basaltic rocks produced at the rift zone between Greenland and the Faroes reach a thickness of 5 km and display velocities of approximately 4.4–5.2 km s^{-1} (Richardson *et al.* 1998). Where the palaeotopography adjacent to the rift was either flat-lying or formed by gently dipping sediments, the lavas were able to flow long distances uninterrupted by topographical barriers (Fliedner & White 2003). For example, east of the Faroe Islands, lava flows extend 150 km away from the islands and cover older sediments (Fliedner & White 2003). The feather edge of these basalt flows is mapped on a number of seismic refraction lines, such as Mobil line 2 (Fig. 5d) (Makris *et al.* 2009) or AMG line 1 (Fig. 5e) (Raum *et al.* 2005). As discussed in Petersen & Funck (2016), the velocity models of the seismic refraction lines in the Faroe–Shetland Basin match fairly well down to the top of the basalt cover, but can show large differences further below. AMG line 1 (Raum *et al.* 2005) terminates close to the Faroe Islands where it has a Moho depth of 22 km (Fig. 5e), which is significantly less than the depths of >30 km observed on the FIRE offshore line (Fig. 5c) (Richardson *et al.* 1998) in an extension of AMG line 1. Receiver function analysis on the Faroe Islands provides an estimate of 29–32 km for the depth to Moho (Harland *et al.* 2009). These values are compatible with what is observed on the Fugloy Ridge (iSIMM Faroes line in Fig. 5d) (Roberts *et al.* 2009), a little to the north of the main axis of the GIFR.

Continental margins north of Iceland

In this section, individual margin segments are described in a clockwise direction starting with the NE Greenland continental margin to the north of the GIFR and finishing with the margins bordering the JMMC. After the review of the margins, five conjugate transects are discussed, one of which is based on gravity inversion (Haase *et al.*, this volume, in press) and not on seismic refraction data, as for the others.

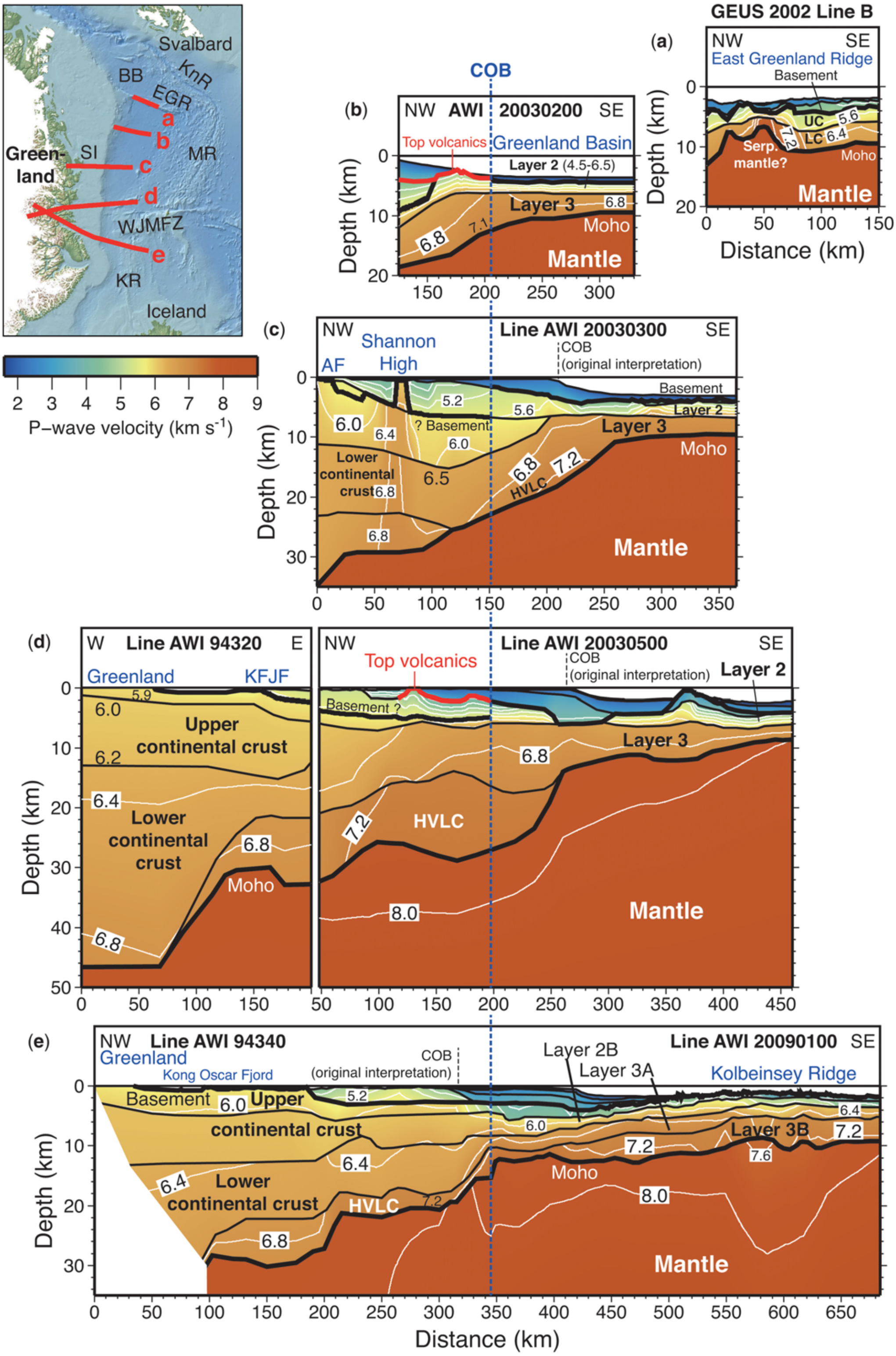
NE Greenland continental margin

The NE Greenland continental margin (Fig. 1) can be divided into three main segments. The southernmost segment comprises the area between the GIFR and the West Jan Mayen Fracture Zone (WJMFZ). The central segment spans the region between the WJMFZ and the East Greenland Ridge, while the northernmost region developed as a shear margin in the De Geer Zone megashear system linking the Atlantic and Arctic spreading systems (Doré *et al.* 2015).

Between the GIFR and the WJMFZ. Gernigon *et al.* (2015) recognized a Mid-Eocene kinematic event at around magnetic Chron C21r (48 Ma) in the Norway Basin that coincides with the onset of dyking and increasing rifting activity between the proto-JMMC and the East Greenland margin. Separation of the JMMC from Greenland started at approximately 30 Ma (Gaina *et al.* 2009). The southern JMMC was completely detached from Greenland by magnetic Chron C6 (20 Ma) (Gaina *et al.* 2009), leading to the accretion of oceanic crust along the Kolbeinsey Ridge.

Hermann & Jokat (2016) published a composite velocity model across this margin segment using lines AWI 94340 and AWI 20090100 (Fig. 6e). The maximum constrained Moho depth on the line

Fig. 5. P-wave velocity models along the Greenland-Iceland-Faroe Ridge and the Faroe-Shetland Basin (FSB). (**a**) SIGMA line 1 (after Holbrook *et al.* 2001); (**b**) ICEMELT line (after Darbyshire *et al.* 1998); (**c**) FIRE offshore line (after Richardson *et al.* 1998); (**d**) iSIMM Faroe line (after Roberts *et al.* 2009), Mobil line 2 (after Makris *et al.* 2009); and (**e**) AMG line 1 (after Raum *et al.* 2005). Abbreviations: COB, continent-ocean boundary; HVLC, high-velocity lower crust; IFR, Iceland-Faroe Ridge.



Svalbard
BB
KnR
EGR
a
b
Green-
land
SI
MR
c
d
WJMFZ
e
KR
Iceland
P–wave velocity (km s-1)
GEUS 2002 Line B
(a)
NW SE
East Greenland Ridge
Basement
UC
LC
Serp. mantle?
Moho
Mantle
Depth (km)
Distance (km)
COB
(b)
NW AWI 20030200 SE
Top volcanics
Greenland Basin
Layer 2 (4.5-6.5)
Layer 3
Moho
Mantle
(c)
NW
Shannon High
AF
Line AWI 20030300 SE
COB (original interpretation)
Basement
Layer 2
? Basement
Layer 3
Lower continental crust
HVLC
Moho
Mantle
(d)
W Line AWI 94320 E
Greenland
KFJF
Upper continental crust
Lower continental crust
Moho
NW Line AWI 20030500 SE
Top volcanics
COB (original interpretation)
Layer 2
Basement ?
Layer 3
HVLC
Mantle
(e)
NW Line AWI 94340
Greenland
Kong Oscar Fjord
COB (original interpretation)
Layer 2B
Layer 3A
Line AWI 20090100 SE
Kolbeinsey Ridge
Basement
Upper continental crust
Lower continental crust
HVLC
Layer 3B
Moho
Mantle
Distance (km)

is 30 km, and the thinning of the continental crust occurs in two steps around km 210 and km 320 in the model. A 150 km-wide and up to 3 km-thick HVLC marks the base of the crust in the COT. The HVLC is relatively poorly constrained (Hermann & Jokat 2016), but its thickness is substantially less than is found in the COT north of the WJMFZ where the high-velocity lower-crustal body displays a thickness of 15 km (Fig. 6d). Hermann & Jokat (2016) argued that the high-velocity body on lines AWI 94340 and 20090100 (Fig. 6e) is the product of excess magma production that was focused along the WJMFZ during the break-up of the JMMC from Greenland. It should be noted that no such HVLC is observed on the conjugate western side of the JMMC (Kodaira *et al.* 1998*b*), which might be expected if the magma production were related to break-up. This could either suggest that the rather poorly constrained HVLC on the Greenland side is not real or that such a layer was missed in the model of the western JMMC. On the Greenland side, there is at least one other line in support for a thin HVLC (Weigel *et al.* 1995), but the wide-angle seismic constraints are also very poor. At the western margin of the JMMC, 4 km-thick continental crust lies adjacent to 9 km-thick oceanic crust (Kodaira *et al.* 1998*a*). The thin continental crust is constrained by only two ocean bottom seismometers (OBSs) 45 km apart from each other, which may not be sufficient to map a thin HVLC.

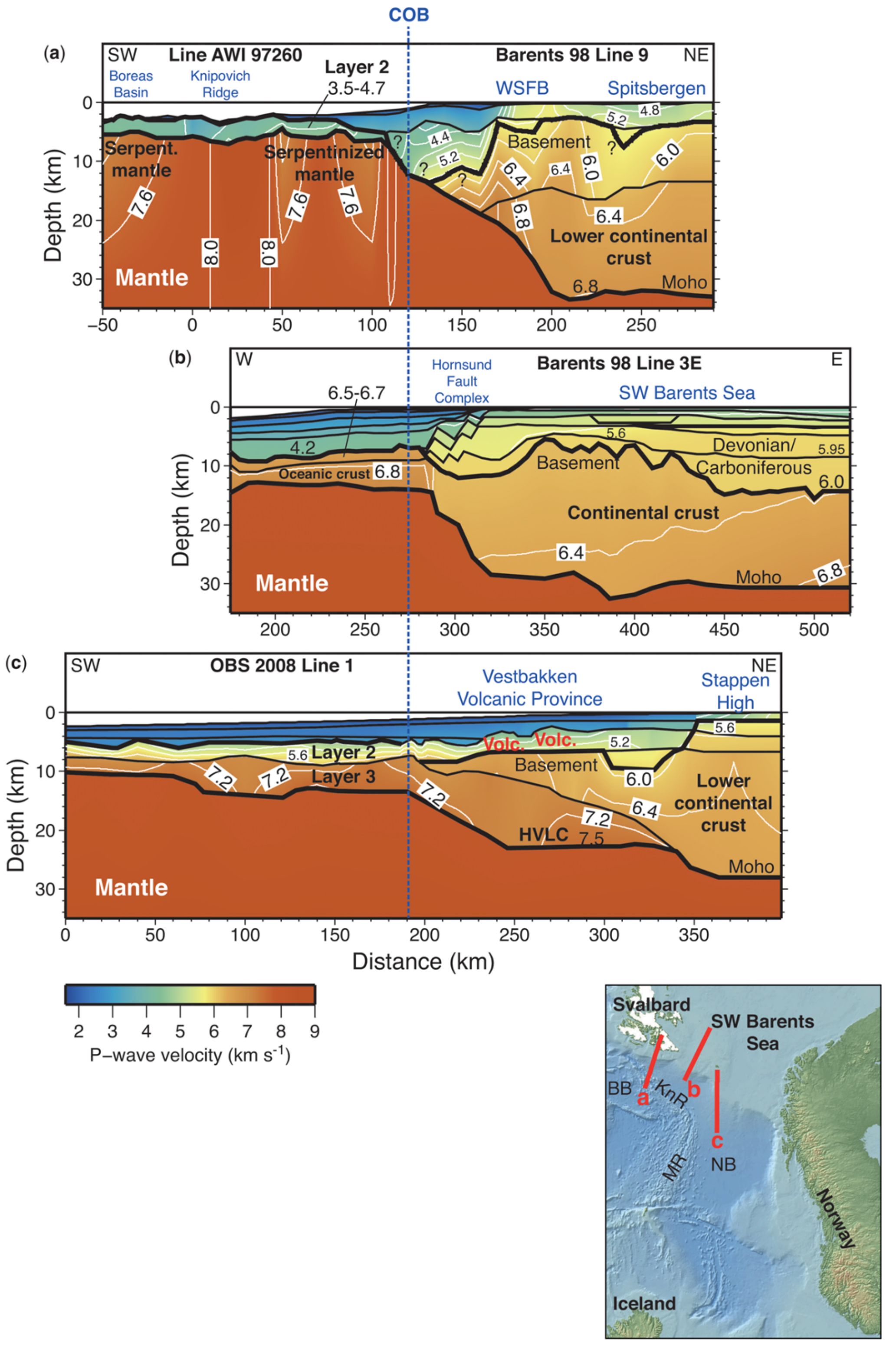
Along the transect AWI 94340 and AWI 20090100 (Fig. 6e), the COB of Funck *et al.* (2014) fits fairly well with the lower-crustal velocity distribution, even though Hermann & Jokat (2016) suggested a location 30 km further landward. The oceanic crust between the COB and the present Kolbeinsey Ridge has a fairly constant thickness of about 9 km.

The sedimentary succession along lines AWI 94340 and 20090100 (Fig. 6e) has a maximum thickness of 3 km (Hermann & Jokat 2016). Thicker sedimentary basins along this margin segment are found onshore Greenland where seismic refraction data indicate a depth of 15 km for the Jameson Land Basin (Weigel *et al.* 1995). Larsen & Marcussen (1992) suggested that the basin infill there might even be up to 18 km thick. Following the post-Caledonian extension with fault-controlled Devonian basins, late Jurassic–early Cretaceous rifting resulted in the deposition of marine sediments over the Devonian sediments in Jameson Land (Surlyk 1990).

Between the WJMFZ and the East Greenland Ridge. The margin segment between the WJMFZ and the East Greenland Ridge is characterized by massive break-up-related Cenozoic volcanism, especially in the southern part. Three seismic refraction transects are shown in Figure 6b–d, illustrating the northwards decrease of magmatic addition to the margin. In the south, on line AWI 20030500 (Fig. 6d), the magmatism is manifested in a 15 km-thick high-velocity lower-crustal body and an up to 5 km-thick series of volcanic rocks at the top of the crust (Voss & Jokat 2007). At the northernmost available line (AWI 20030200: Fig. 6b), hardly any signs of break-up-related volcanism are left. Maximum velocities in the lower crust close to the COB are 7.1 km s^{-1} and the total crustal thickness there is only 7 km (Voss *et al.* 2009). Landward of the COB, seismic reflection data indicate the presence of basalts forming an outer high (Voss *et al.* 2009). Line AWI 20030200 does not image the proximal part of the margin due to sea ice that inhibited the seismic data acquisition there.

Line AWI 20030300 (Fig. 6c) at the centre of the margin segment displays a more pronounced HVLC than the line discussed previously. In the continental domain, Voss *et al.* (2009) noticed a positive velocity anomaly beneath the Shannon High. East of the high, the thickness of the basalts and the depth to basement are poorly resolved due to a lack of velocity contrasts. The problem is that the velocities in the deeper Mesozoic and Palaeozoic sediments are similar to the ones expected in crystalline crust, which makes it difficult to distinguish between the two. Voss *et al.* (2009) suggested that the rift-related basin infill might be up to 15 km thick. In Figure 6c, the basement is indicated by the 5.7 km s^{-1} contour, similar to what is done along the lines in the north and south (Fig. 6b, d). This leaves a substantial uncertainty in the thickness of the rift-related basins along this margin segment. In the northern part of the margin, no seismic refraction data are available. However, seismic reflection data from the KANUMAS group released in 2014 can be used to compile a sediment thickness map for the NE Greenland Shelf (Hopper *et al.*, this

Fig. 6. P-wave velocity models at the NE Greenland margin: (**a**) GEUS2002 line B (after Døssing & Funck 2012); (**b**) line AWI 20030200 (after Voss *et al.* 2009); (**c**) line AWI 20030300 (after Voss *et al.* 2009); (**d**) lines AWI 94320 (after Schlindwein & Jokat 1999) and AWI 20030500 (after Voss & Jokat 2007); and (**e**) lines AWI 94340 and 20090100 (after Hermann & Jokat 2016). Abbreviations: AF, Ardencaple Fjord; BB, Boreas Basin; COB, continent–ocean boundary; EGR, East Greenland Ridge; HVLC, high-velocity lower crust; KFJF, Kejser Franz Joseph Fjord; LC, lower crust; Serp., Serpentinized; SI, Shannon Island; UC, upper crust; WJMFZ, West Jan Mayen Fracture Zone.


COB
(a)
SW
Line AWI 97260
Boreas Basin
Knipovich Ridge
Layer 2
3.5-4.7
Barents 98 Line 9
NE
WSFB
Spitsbergen
Serpent. mantle
Serpentinized mantle
Basement
Lower continental crust
Mantle
Moho
Depth (km)
(b)
W
Barents 98 Line 3E
E
Hornsund Fault Complex
SW Barents Sea
6.5-6.7
Oceanic crust
Basement
Devonian/ Carboniferous
Continental crust
Mantle
Moho
Depth (km)
(c)
SW
OBS 2008 Line 1
NE
Vestbakken Volcanic Province
Stappen High
Layer 2
Layer 3
Volc.
Volc.
Basement
Lower continental crust
HVLC
Mantle
Moho
Depth (km)
Distance (km)
P–wave velocity (km s-1)
Svalbard
SW Barents Sea
BB
KnR
MR
NB
Norway
Iceland

volume, in prep). This map indicates less than 5 km of sediments to the east of Shannon Island. Thicker sedimentary basins are observed in the north, where the sedimentary pile is up to 17 km thick in the North Danmarkshavn Basin.

Another point of interest is the location of the COB that is subject to some controversy. In the original publications, with the velocity models of lines AWI 20030300 and 20030500 (Fig. 6c, d), the COB is located approximately 60 km seaward of the one proposed by Funck *et al.* (2014) that is, to a large degree, determined by plate reconstructions (Gaina 2014; cf. Gaina *et al.*, this volume, in review) using the arguably better-defined location of the COB at the conjugate mid-Norwegian margin. The originally proposed COB along line AWI 20030300 (Fig. 6c) (Voss *et al.* 2009) fits well with the modelled lower-crustal velocity variations, while the velocities on line AWI 20030500 (Fig. 6d) would, instead, support the more landward COB of Funck *et al.* (2014).

North of the East Greenland Ridge. The margin north of the East Greenland Ridge (EGR) developed as a shear margin and is the least-studied margin in the NE Atlantic owing to the year-round cover with sea ice. The continental domain of the margin is not studied with seismic refraction lines, with the exception of the EGR that developed along the Greenland Fracture Zone and protrudes from the NE Greenland shelf into the oceanic basins (Fig. 1). The ridge (Fig. 6a) is a 250 km-long and up to 50 km-wide bathymetric high that is composed of continental crust thinned to 2–6 km and locally underlain by partially serpentinized mantle (Døssing *et al.* 2008; Døssing & Funck 2012; Gerlings *et al.* 2014; Funck *et al.* 2015). The EGR was sheared from the shelf and has a complex internal structure with two overstepping main ridge segments (Døssing & Funck 2012).

Immediately to the north of the EGR in the SW Boreas Basin, seismic refraction and coincident reflection data indicate the presence of an extremely thin and faulted transitional crust (Døssing *et al.* 2008). The exact extent of this crust is unknown, but can be correlated at least 40 km to the north of the central part of the EGR (Døssing *et al.* 2008). Discontinuous and often weak magnetic lineations characterize the Boreas Basin (cf. Engen *et al.* 2008). Hermann & Jokat (2013) showed the presence of thin (generally <3 km) oceanic crust in the basin.

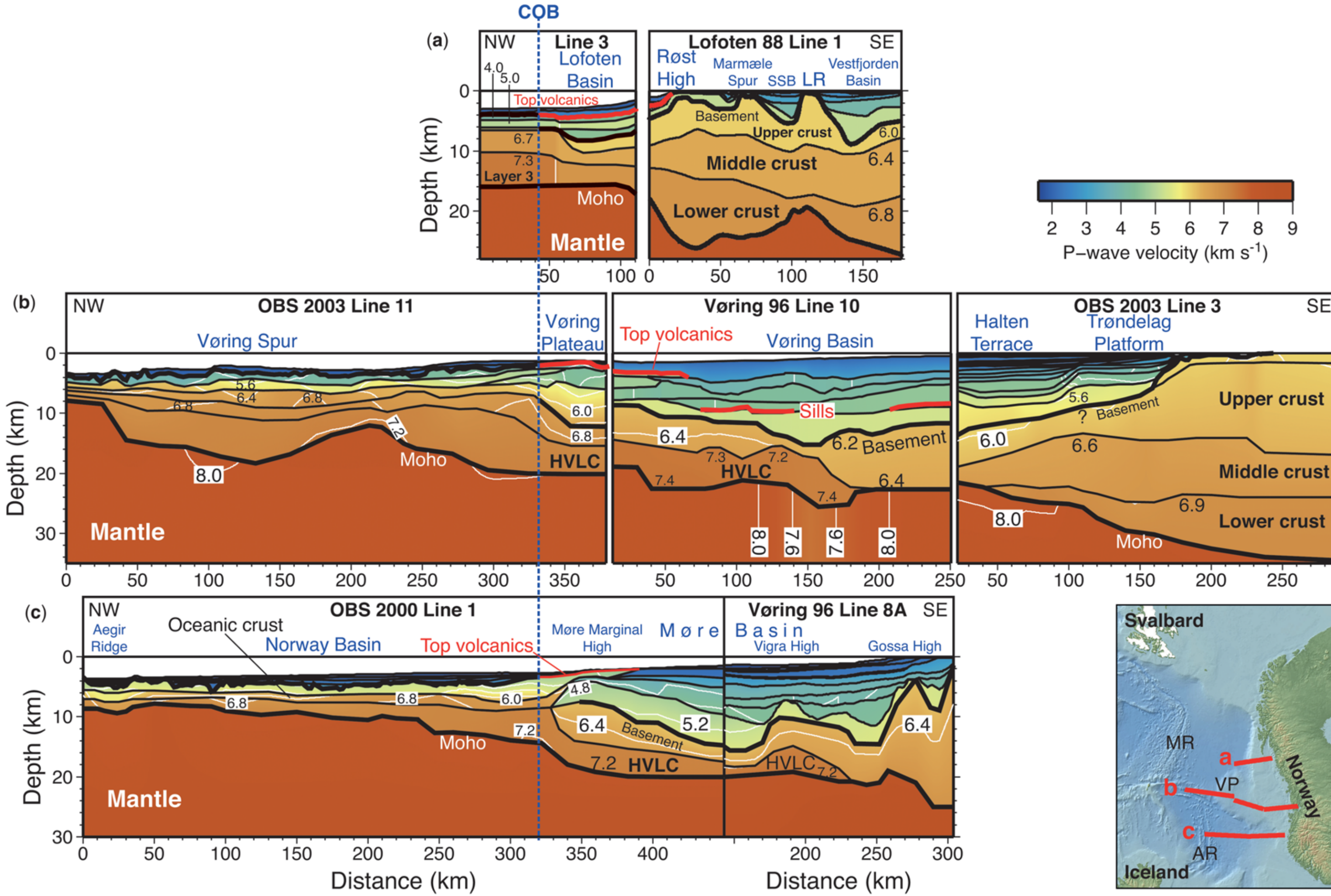
SW Barents Sea margin

The SW Barents Sea margin is a transform margin that extends from northern Norway to Svalbard (Fig. 1). The Barents Sea is a wide rifted continental shelf domain composed of numerous fault-bounded interconnected and segmented basins that are linked both to the Atlantic and Arctic rift systems. The formation of the western Barents Sea began in the Carboniferous, followed by two additional rift phases in Middle Jurassic–Early Cretaceous times and in the early Cenozoic (Faleide *et al.* 1993, 2008). The basins in the Barents Sea locally exceed 18 km in depth (Ritzmann *et al.* 2007).

Late Cretaceous rifting, subsequent break-up and initial seafloor spreading in the Norwegian–Greenland Sea was linked to the Eurasia Basin in the Arctic by the regional De Geer Zone mega-shear system (Faleide *et al.* 2008). The southern limit of this predominantly sheared margin system in the SW Barents Sea is marked by the Senja Fracture Zone, the conjugate to the Greenland Fracture Zone. The margin is divided into two large shear segments, the Hornsund and Senja margins (Fig. 1), and a central rifted segment with volcanism, the Vestbakken margin. At Svalbard, initial shearing was followed by rifting, while north of Svalbard and up to the Yermak Plateau, the margin developed as a complex sheared and rifted margin (Faleide *et al.* 2008).

Line 3E of the Barents 98 survey (Breivik *et al.* 2003) is a dip line at the Hornsund margin (Fig. 7b) and illustrates the shear-margin setting. The Moho shallows from 29 to 14 km over a distance of only 35 km and the transition from continental to oceanic crust can be resolved to lie within a narrow zone of less than 5 km in width (Breivik *et al.* 2003). The upper part of the continental crust in the vicinity of the Hornsund fault complex is dominated by two large, rotated downfaulted blocks with throws of 2–3 km on each fault, apparently formed during the transform margin development (Breivik *et al.* 2003). Beneath the shelf, the top of the crystalline basement is as deep as 16 km and is primarily constrained by a coincident seismic reflection line. Velocities in the interpreted Devonian–Carboniferous sedimentary section vary between 5.6 and 6.0 km s^{-1}, which would be difficult to distinguish from crystalline basement using seismic refraction data alone. The oceanic crust adjacent to the COB has a thickness of 7 km, but thins to as little as 4 km further to the west. The oceanic crust lacks

Fig. 7. P-wave velocity models at the SW Barents Sea margin: (**a**) lines AWI 97260 and Barents 98 line 9 (after Ritzmann *et al.* 2002); (**b**) Barents 98 line 3E (after Breivik *et al.* 2003); and (**c**) line 1 of the OBS 2008 survey (Libak *et al.* 2012). Abbreviations: BB, Boreas Basin; COB, continent–ocean boundary; HVLC, high-velocity lower crust; KnR, Knipovich Ridge; MR, Mohns Ridge; NB, Norway Basin; Serpent., Serpentinized; Volc., Volcano; WSFB, West Spitsbergen Fold Belt.



COB
(a)
NW
Line 3
Lofoten
Basin
4.0
5.0
Top volcanics
6.7
7.3
Layer 3
Moho
Mantle
Depth (km)
0
10
20
50
100
Lofoten 88 Line 1
SE
Røst
High
Marmæle
Spur
SSB
LR
Vestfjorden
Basin
Basement
Upper crust
6.0
Middle crust
6.4
Lower crust
6.8
0
50
100
150
2
3
4
5
6
7
8
9
P–wave velocity (km s-1)
(b)
NW
OBS 2003 Line 11
Vøring Spur
Vøring
Plateau
5.6
6.4
6.8
6.8
6.8
7.2
6.0
6.8
HVLC
Moho
8.0
Mantle
Depth (km)
0
10
20
30
0
50
100
150
200
250
300
350
Vøring 96 Line 10
Top volcanics
Vøring Basin
Sills
6.4
6.2
Basement
7.3
7.2
HVLC
7.4
6.4
7.4
8.0
7.6
7.6
8.0
50
100
150
200
250
OBS 2003 Line 3
SE
Halten
Terrace
Trøndelag
Platform
5.6
?
Basement
Upper crust
6.0
6.6
Middle crust
6.9
Lower crust
8.0
Moho
50
100
150
200
250
(c)
NW
OBS 2000 Line 1
Oceanic crust
Aegir
Ridge
Norway Basin
Top volcanics
Møre Marginal
High
M ø r e
B a s i n
6.8
6.8
6.0
4.8
7.2
6.4
Basement
5.2
Moho
7.2
HVLC
Mantle
Depth (km)
0
10
20
30
0
50
100
150
200
250
300
350
400
Distance (km)
Vøring 96 Line 8A
SE
Vigra High
Gossa High
6.4
HVLC
7.2
200
250
300
Distance (km)
Svalbard
MR
a
VP
b
Norway
c
AR
Iceland

velocities typical for oceanic layer 2, but is characterized by basement velocities of 6.6 km s^{-1}. Breivik *et al.* (2003) saw this as a response to mineral infilling and the closure of cracks, fissures and voids in layer 2 induced by the 7–8 km-thick sedimentary overburden.

The rifted Vestbakken margin segment is illustrated by line 1 of the OBS 2008 experiment (Fig. 7c) (Libak *et al.* 2012). In contrast to the rapid crustal thinning observed at the Hornsund margin, the shallowing of the Moho is more gradual from a depth of 28 to 13 km over a 170 km-wide zone. Landward of the COB, increased velocities of up to 7.5 km s^{-1} are observed in the lower crust, which are interpreted as intrusions (Libak *et al.* 2012). Within the region of the HVLC, volcanoes are observed (their position is indicated in Fig. 7c) and Libak *et al.* (2012) interpreted the upwarping of velocity contours in the lowermost sedimentary layer (velocities of 4.4–5.2 km s^{-1}) as marking feeder dykes. Prominent volcanoes, as well as sill intrusions, are interpreted in seismic reflection data at the outer margin at the Vestbakken Volcanic Province with two phases of volcanism in the early Eocene and early Oligocene (Faleide *et al.* 1988). The oceanic crust along this profile has a general thickness of 7 km, with variations of between 5 and 9 km. Velocities in the interpreted oceanic layer 3 (6.8–7.7 km s^{-1}) are rather high and could also be compatible with partially serpentinized mantle, a possibility that was not ruled out by Libak *et al.* (2012). In this case, the thickness of the oceanic crust would be overestimated.

The Hornsund margin is presented by a seismic refraction transect (Fig. 7a) that uses data from two experiments (line AWI 97260 and Barents 98 line 9). On this profile, the oceanic crust is characterized by a 3.5 km-thick layer 2 with velocities of between 3.5 and 5.7 km s^{-1}, while a layer 3 could not be mapped (Ritzmann *et al.* 2002). Velocities of 7.3–8.0 km s^{-1} beneath the oceanic crust indicate a partial serpentinization of the mantle rocks. The thinning of the continental crust occurs over an 80 km-wide zone extending from the West Spitsbergen Fold Belt seaward. In this zone, the Moho shallows from 33 to 7 km. Between model km 150 and 180, a region with increased lower-crustal velocities (>7.2 km s^{-1}) is observed, although the seismic resolution is low. Ritzmann *et al.* (2002) related these velocities to magmatic intrusions of unknown origin. The nature of the deep sedimentary basin landward of the COB is not clear and the depth of the crystalline basement is not resolved by the data. The crystalline crust beneath Spitsbergen has a thickness of 30 km, with velocities ranging from 5.5 to 6.8 km s^{-1} and is overlain by a 3–7 km-thick sedimentary sequence.

Mid-Norwegian continental margin

The mid-Norwegian continental margin is divided into three main segments, referred to as the Møre, Vøring, and Lofoten margins (Fig. 1). These segments are delimited by regional transfer zones. The Vøring margin is bounded by the Bivrost and Jan Mayen lineaments to the north and south, respectively. At the Jan Mayen Lineament, Olesen *et al.* (2007) notice only a weak expression in basement structure while the related East Jan Mayen Fracture Zone is associated with a major shift in the COB. In contrast, the Bivrost Lineament is well expressed in the basement structure but lacks an outboard fracture zone according to the interpretation of Olesen *et al.* (2007), who relate a previously proposed Bivrost Fracture Zone to data artefacts. Doré *et al.* (1997) suggest that the transfer zones are likely linked to the structural heritage of the region.

A first extensional phase took place during Devonian–Carboniferous times and was possibly related to the gravitational collapse of the Caledonian orogen (Seranne & Seguret 1987; Andersen & Jamtveit 1990; Osmundsen & Andersen 2001 and references therein). Early rift basins formed most likely during Carboniferous, Permian and Middle Triassic times in the proximal setting (e.g. the Trøndelag Platform), while renewed extension in the Middle Jurassic and in the Late Jurassic to Early/middle-Cretaceous shaped the distal basins (e.g. the Lofoten, Vøring, and Møre basins) (Doré *et al.* 1999; Brekke 2000). The complex outer ridges were affected by the last rifting phase of Late Cretaceous–Paleocene age (Gernigon *et al.* 2003; Ren *et al.* 2003).

The mid-Norwegian margin is densely sampled by seismic refraction lines (see figure 2 in Funck *et al.* 2016) but individual profiles are rather short and do not cover the entire margin from the proximal to the distal domain. Hence, composite profiles were constructed to illustrate the crustal structure of the three margin segments (Fig. 8).

Fig. 8. P-wave velocity models at the mid-Norwegian margins: (**a**) the Lofoten margin with lines 1 (after Mjelde *et al.* 1993) and 3 (after Mjelde *et al.* 1992) of the Lofoten 88 survey; (**b**) the Vøring margin with lines 3 (after Breivik *et al.* 2011) and 11 (after Breivik *et al.* 2008) of the OBS 2003 survey and Vøring 96 line 10 (after Raum *et al.* 2002); and (**c**) the Møre margin with line 1 of the OBS 2000 survey (after Breivik *et al.* 2006) and Vøring 96 line 8A (after Raum 2000). Abbreviations: AR, Aegir Ridge; COB, continent–ocean boundary; HVLC, high-velocity lower crust; LR, Lofoten Ridge; MR, Mohns Ridge; SSB, Skomvær Sub-basin; VP, Vøring Plateau.

Lofoten margin. The transect across the Lofoten margin is composed of lines 1 and 3 (Fig. 8a) of the Lofoten 88 experiment (Mjelde *et al.* 1992, 1993). The imaged continental crust has a maximum thickness of 25 km beneath the Røst High. Landward of the high, the transect crosses the Marmæle Spur, the Lofoten Ridge and the Vestfjorden Basin, resulting in a variable basement depth of between 1 and 9 km that is consistent with a highly faulted basement, as indicated in the interpretation of the coincident seismic reflection data (Mjelde *et al.* 1993). Volcanics extend from the western flank of the Røst High onto the oceanic crust. The seismic resolution in the COT is limited as the spacing of OBSs was rather high (*c.* 30 km). The velocity model indicates a sharp transition between the continental and oceanic crust, which is different from modern experiments that observe a smoother velocity transition in the COT of magma-rich margins (e.g. White & Smith 2009). This may explain why the COB as interpreted by Funck *et al.* (2014) is further seaward than the velocity model of Mjelde *et al.* (1993) may suggest. In any case, velocities in the lower oceanic crust are modelled at 7.3 km s^{-1}, which is compatible with high-velocity lower crust commonly observed at magma-rich margins. However, landward of the COB, no distinct HVLC is mapped. This distinguishes the Lofoten margin from the Vøring and Møre margins to the south. The thickness of the initial oceanic crust is 12 km, including the overlying flood basalts.

Vøring margin. The Vøring margin is much wider than the Lofoten margin. The chosen transect (Fig. 8b) is composed of lines 3 and 11 of the OBS 2003 experiment (Breivik *et al.* 2008, 2011) and line 10 of the Vøring 96 survey (Raum *et al.* 2002). There are some slight deviations in the models where the lines join, which is of no concern here as we look at large-scale structures. At the landward end of the transect, the maximum Moho depth constrained by the data is 32 km. The basement deepens seaward to a depth of 13 km beneath the Halten Terrace. However, beneath the terrace and the Trøndelag Platform there is some uncertainty about the exact location of the basement, as velocities of 5.3–5.7 km s^{-1} are observed that are intermediate between typical crystalline crust and Mesozoic sedimentary strata (Breivik *et al.* 2011).

In the Vøring Basin, the sedimentary section is up to 14 km thick (Fig. 8b), including sills and inner flows (Raum *et al.* 2002). The basin is underlain by thinned continental crust with a thickness of 2–11 km, not including the HVLC (7.2–7.4 km s^{-1}) beneath the western part of the basin that has a maximum thickness of 8 km. Raum *et al.* (2002) interpreted the HVLC as a magmatic underplated body. However, this interpretation is controversial. Under the Rån Ridge at the outer Vøring margin, velocities locally exceed 8.5 km s^{-1} and have been mapped in a tectonically complex setting where densities of 3500 kg m^{-3} are used for gravity modelling (Raum *et al.* 2006). This HVLC is accordingly explained as eclogite because the velocities are too high for both crustal crystalline rocks and mantle peridotite. Other studies use much lower densities for gravity modelling (Rouzo *et al.* 2006; Reynisson *et al.* 2010) and the 3D complexity of the area might suggest that 2D seismic refraction profiles did not adequately map the deep crustal structures in sufficient detail. In contrast, velocities between 7.0 and 7.8 km s^{-1} could represent both mafic intrusions (including magmatic underplating) and partially serpentinized mantle peridotite (Miller & Christensen 1997; Christensen 2004). Lundin & Doré (2011) proposed that the thinned crust beneath the Vøring Basin presents hyper-extended crust that is underlain by partially serpentinized upper mantle. Mjelde *et al.* (2009) favoured a model that relates the HVLC to mafic intrusions, as there is a spatial relationship between the HVLC and sill intrusions. Gernigon *et al.* (2003, 2004) showed, based on high-resolution reflection seismic data and gravity modelling, that the HVLCs of the outer Vøring margin correlate with structural highs and that crustal thinning occurred adjacent to them. Thus, the HVLCs were already in place prior to break-up and probably represent high-pressure metamorphic rocks and not serpentinized mantle.

Beneath the Vøring Plateau, crustal velocities increase seaward up to the position of the COB (Fig. 8b). At the onset of seafloor spreading, the oceanic crustal thickness is 18 km, but it thins to 9 km over a distance of 115 km. However, beneath the Vøring Spur, a renewed crustal thickening is observed to a maximum of 15 km. Breivik *et al.* (2008) relate this to magmatic underplating beneath oceanic crust.

Møre margin. Compared to the Vøring margin, the Møre margin is significantly narrower. The transect shown in Figure 8c is composed of line 1 of the OBS 2000 experiment in the NW (Breivik *et al.* 2006), while the SE part is taken from line 8A of the Vøring 96 survey (Raum 2000). The full-thickness continental crust is not really covered by this transect as the shelf is relatively narrow and no land stations were deployed. However, other lines nearby obtain Moho depths of 29 km beneath the shelf and 37 km onshore (Kvarven *et al.* 2014). Faulted continental crust is observed inboard of the Gossa High, and HVLC has been mapped here in a number of seismic reflection and refraction studies (Olafsson *et al.* 1992; Kvarven *et al.* 2014; Nirrengarten

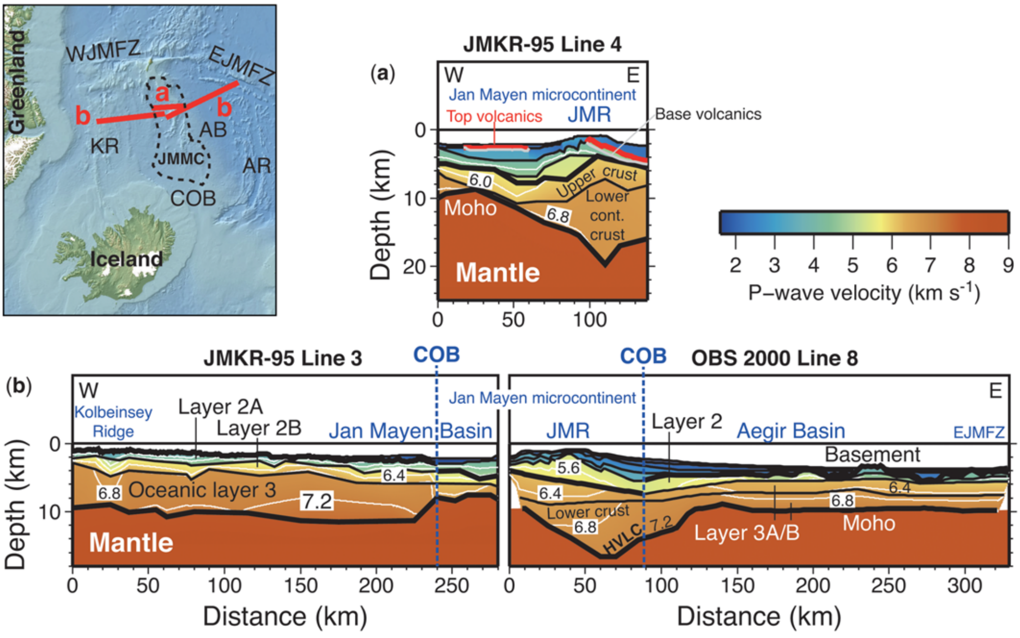


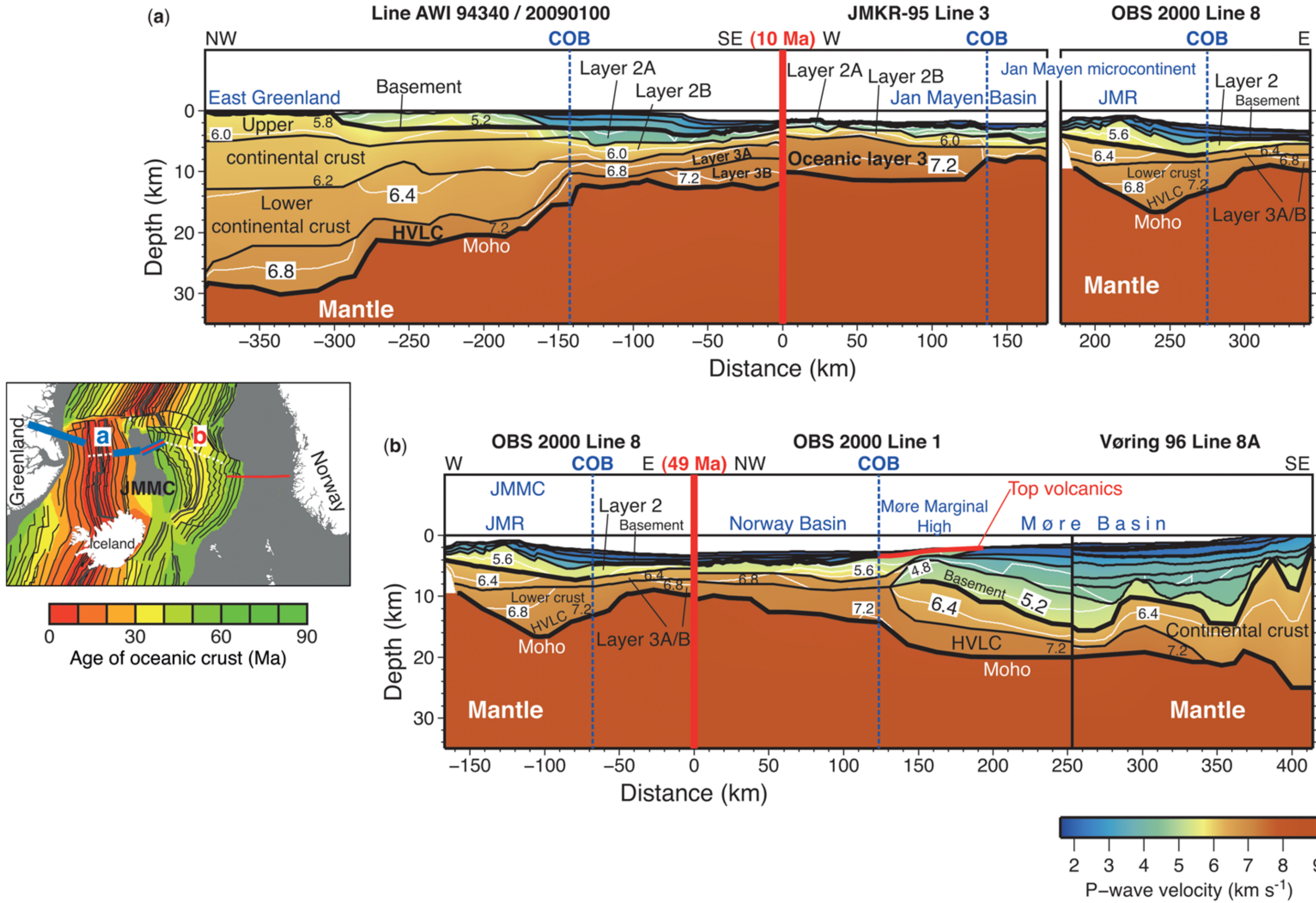
Fig. 9. P-wave velocity models at the Jan Mayen microcontinent: (**a**) JMKR-95 line 4 (after Kodaira *et al.* 1998*b*); and (**b**) JMKR-95 line 3 (after Kodaira *et al.* 1998*a*) and line 8 of the OBS 2000 survey (after Mjelde *et al.* 2008). Abbreviations: AB, Aegir Basin; AR, Aegir Ridge; COB, continent–ocean boundary; cont., continental; EJMFZ, East Jan Mayen Fracture Zone; HVLC, high-velocity lower crust; JMMC, Jan Mayen microcontinent; JMR, Jan Mayen Ridge; KR, Kolbeinsey Ridge; WJMFZ, West Jan Mayen Fracture Zone.

et al. 2014), as well as inferred from isostatic modelling (Gradmann *et al.*, this volume, in review). Seaward, the continental crust is thinned to 3–7 km, overlain by the up to 14 km-thick sedimentary sequence of the Møre Basin and underlain by an up to 4 km-thick HVLC with velocities of around 7.2 km s^{-1}. Similar to the Vøring margin, the nature of the HVLC is also under debate at the Møre margin. Lundin & Doré (2011) again advocated partially serpentinized mantle peridotite. Kvarven *et al.* (2014) noticed velocity variations within the HVLC, with higher velocities in the west (7.6–7.7 km s^{-1}) than in the east (7.2 km s^{-1}), although not very well constrained. They proposed that the higher velocities could indicate the presence of a partially eclogized body, while the lower velocities are interpreted as magmatic underplating. In a recent review of the HVLC beneath the Møre margin, Nirrengarten *et al.* (2014) concluded that the HVLC in the proximal part of the margin is most likely to represent inherited crustal bodies and not rift-related serpentinized mantle. For the distal part of the margin, their preferred interpretation is that the HVLC is made of boudins of hyper-extended, pre-rift lower-continental crustal rocks more or less intruded by Early Tertiary magmatic material.

The Møre Marginal High is covered by an up to 4 km-thick volcanic sequence (Fig. 8c). The initial oceanic crustal thickness is 10 km, which is similar to the Lofoten margin but substantially less than at the Vøring margin (18 km). Further seaward, a relatively homogeneous oceanic crust, with a thickness of 5–7 km, is observed beneath the Norway Basin.

Margins of the Jan Mayen microcontinent

The Jan Mayen microcontinent (JMMC) is a fragment of continental crust extending from north of Iceland up to the East Jan Mayen Fracture Zone (EJMFZ: Fig. 1). Jan Mayen Island itself is not considered part of the JMMC as geochemical data provide no evidence of continental contamination of the rocks on the island (Svellingen & Pedersen 2003). Gaina *et al.* (2009) proposed that the major Oligocene plate-boundary reorganization and microcontinent formation might have been preceded by various ridge propagations and/or short-lived triple junctions NE and, possibly, SW of the JMMC from the initiation of seafloor spreading (54 Ma) to Chron C18 (40 Ma). This resulted in the formation of a highly extended or even fragmented JMMC, and subsequent deformation of its margins and surrounding regions.


(a)
Line AWI 94340 / 20090100
JMKR-95 Line 3
OBS 2000 Line 8
NW
COB
SE (10 Ma) W
COB
COB
E
East Greenland
Basement
Layer 2A
Layer 2B
Layer 2A
Layer 2B
Jan Mayen Basin
Jan Mayen microcontinent
JMR
Layer 2
Basement
Upper
continental crust
Lower
continental crust
HVLC
Moho
Mantle
Layer 3A
Layer 3B
Oceanic layer 3
Lower crust
Layer 3A/B
Depth (km)
Distance (km)
(b)
OBS 2000 Line 8
OBS 2000 Line 1
Vøring 96 Line 8A
W
COB
E (49 Ma) NW
COB
SE
JMMC
JMR
Layer 2
Basement
Norway Basin
Møre Marginal High
Top volcanics
Møre Basin
Lower crust
HVLC
Moho
Layer 3A/B
Continental crust
Mantle
Depth (km)
Distance (km)
Greenland
a
b
JMMC
Iceland
Norway
Age of oceanic crust (Ma)
P–wave velocity (km s^{-1})

During Paleocene rifting, the JMMC was still part of Greenland (Gaina *et al.* 2009). Break-up in the Early Eocene occurred between the JMMC and the mid-Norway and Faroe margins along the Aegir Ridge, and is imaged by regional SDRs along the eastern margin of the JMMC (Peron-Pinvidic *et al.* 2012). At 30 Ma, the Aegir Ridge became extinct and the separation of the JMMC from East Greenland was completed at 20 Ma (Gaina *et al.* 2009) when a seafloor spreading system developed along the Kolbeinsey Ridge. Gaina *et al.* (2009) suggested that the southernmost extended, fragmented character of the southernmost JMMC is a product of several failed ridge-propagation attempts of the Kolbeinsey Ridge. This is also the reason for some uncertainty in defining the COB in the southern JMMC (cf. Funck *et al.* 2014).

The combination of JMKR-95 line 3 (Kodaira *et al.* 1998*a*) and OBS 2000 line 8 (Mjelde *et al.* 2008) presents a seismic refraction transect from the Kolbeinsey Ridge across the JMMC and into the Aegir Basin (Fig. 9b). The western and eastern margins of the JMMC look distinctively different. In the west, a 40 km-wide zone with thin continental crust (thickness of 2.5–4 km) is observed with no indication of a HVLC. In contrast, the eastern margin is characterized by up to 10 km-thick crystalline crust with a high-velocity lower-crustal body (up to 7.2 km s^{-1}) at the base. A nearby line (JMKR-95 line 4: Fig. 9a) indicates the presence of basalts extending from the top of the Jan Mayen Ridge eastwards (Kodaira *et al.* 1998*b*). This line also displays a layer with velocities of 4.6–5.0 km s^{-1} in the Jan Mayen Basin interpreted as basalts that were erupted at 30 Ma (Kodaira *et al.* 1998*b*). In the area of the northern JMMC, the sedimentary column has a maximum thickness of 6 km observed on the western flank of the Jan Mayen Ridge (Fig. 9a).

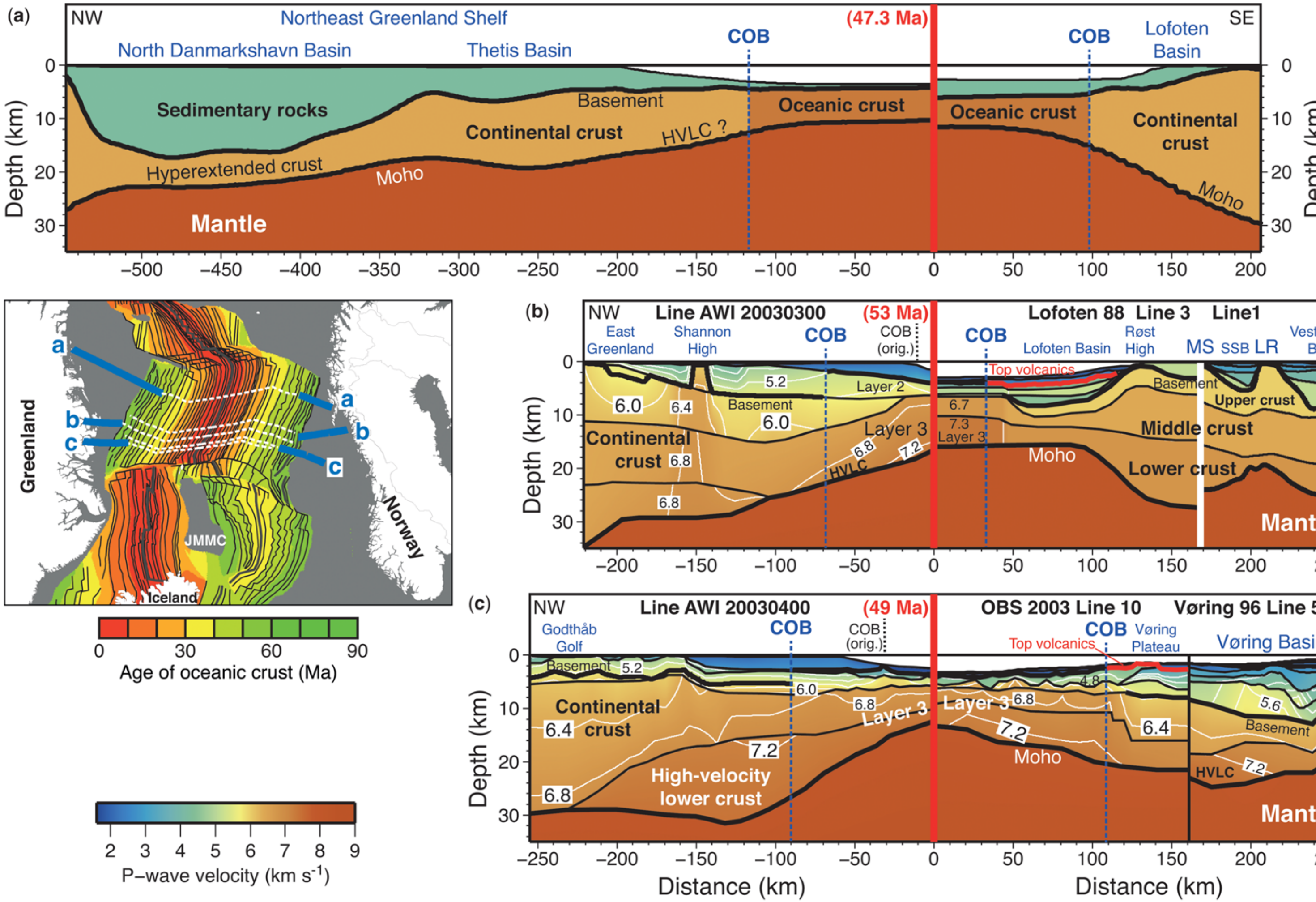
Moho depth beneath the JMMC is 17–19 km on the lines shown in Figure 9, but close to the northern limit of the JMMC; Kandilarov *et al.* (2012) reported a depth of 27 km. The initial oceanic crust at the eastern margin of the JMMC has a thickness of 9 km, but thins to 5 km over a distance of 30 km (Fig. 9b). Further eastwards, the oceanic crustal thickness varies between 5 and 6 km. The oceanic crust at the western COB of the JMMC has a thickness of 5 km, but increases to 9 km just 15 km to the east. The remainder of JMKR-95 line 3 displays a fairly constant and unusually high oceanic crustal thickness of 9 km.

Conjugate transects north of Iceland

In this section, conjugate transects involving the JMMC are presented first, followed by three transects north of the Jan Mayen Fracture Zone. The western flank of the JMMC (Fig. 10a) is constrained by JMKR-95 line 3 (Kodaira *et al.* 1998*a*) and OBS 2000 line 8 (Mjelde *et al.* 2008). On the Greenland side, line ARK 1988-3 (Weigel *et al.* 1995) is perfectly conjugate to these two lines, but has a rather poor resolution owing to a curved line geometry and a limited number of receivers in the offshore segment of the line. Hence, lines AWI 94340 and 20090100 (Hermann & Jokat 2016) are chosen instead to illustrate this conjugate margin pair (Fig. 10a). There is an offset of about 85 km between the opposite lines. In the oceanic domain, the velocity structure is fairly similar on either side, with 9–10 km of crust. A difference is the division of the lower oceanic crust into a layer 3A and 3B on the Greenland side, while it is modelled as a single layer on the JMMC side.

While the Moho deepens immediately landward of the COB on the Greenland side, the Moho stays almost horizontal for another 40 km at the JMMC before crustal thickening is observed (Fig. 10a). Another difference is the presence of a 2–3 km-thick high-velocity body at the base of the East Greenland crust, while such a lower-crustal body is not indicated at the western margin of the JMMC. As mentioned earlier, the seismic constraints in the HVLC are limited off East Greenland and there is also the possibility that a HVLC at the western margin of the JMMC was missed in the dataset. Hence, the asymmetry with respect to the HVLC may just be perceived. Given the thicker than normal oceanic crust that formed along the Kolbeinsey Ridge, the system seems to be magma-rich rather than magma-poor. If both velocity models are correct, the asymmetry may relate to the focusing of magma production along the WJMFZ during break-up of the JMMC from Greenland, as suggested by Hermann & Jokat (2016). The absence of the HVLC on the JMMC profile can then be attributed to its

Fig. 10. P-wave velocity models of conjugate transects at the JMMC: (**a**) Combined lines AWI 94340 and 20090100 (after Hermann & Jokat 2016), JMKR-95 line 3 (after Kodaira *et al.* 1998*a*), and line 8 of the OBS 2000 survey (after Mjelde *et al.* 2008); and (**b**) lines 8 (after Mjelde *et al.* 2008) and 1 (Breivik *et al.* 2006) of the OBS 2000 survey, and Vøring 96 line 8a (after Raum 2000). The inset map displays the age of the oceanic crust (after Gaina 2014; cf. Gaina *et al.*, this volume, in review), together with the line locations (red and blue lines) and flow lines (dashed white lines). Thin solid lines mark selected isochrons. Abbreviations: COB, continent–ocean boundary; HVLC, high-velocity lower crust; JMMC, Jan Mayen microcontinent; JMR, Jan Mayen Ridge.


(a)
NW
SE
Northeast Greenland Shelf
North Danmarkshavn Basin
Thetis Basin
COB
(47.3 Ma)
COB
Lofoten Basin
Sedimentary rocks
Basement
Oceanic crust
Oceanic crust
Continental crust
Continental crust
HVLC ?
Hyperextended crust
Moho
Moho
Mantle
Depth (km)
Depth (km)
Greenland
Norway
JMMC
Iceland
a
b
c
a
b
c
Age of oceanic crust (Ma)
0
30
60
90
P–wave velocity (km s-1)
(b)
NW
Line AWI 20030300
(53 Ma)
Lofoten 88 Line 3
Line1
SE
East Greenland
Shannon High
COB
COB (orig.)
COB
Lofoten Basin
Røst High
MS SSB LR
Vestfjorden Basin
Top volcanics
Layer 2
Basement
Basement
Upper crust
Layer 3
Layer 3
Continental crust
HVLC
Middle crust
Lower crust
Moho
Mantle
(c)
NW
Line AWI 20030400
(49 Ma)
OBS 2003 Line 10
Vøring 96 Line 5
SE
Godthåb Golf
COB
COB (orig.)
Top volcanics
COB
Vøring Plateau
Vøring Basin
Basement
Continental crust
Layer 3
Layer 3
Basement
High-velocity lower crust
Moho
HVLC
Mantle
Distance (km)
Distance (km)

more distal position to the WJMFZ when compared to the East Greenland line.

At the western margin of the JMMC, the continental crust is only 3–4 km thick in a 40 km-wide zone adjacent to the COB (Fig. 10a). Mantle velocities beneath this hyper-extended crust are 8.0 km s^{-1} (Kodaira *et al.* 1998*a*) and, hence, do not support the presence of partially serpentinized mantle rock. This is, to some degree, surprising as the entire crust becomes brittle at stretching factors of 3–5 (Pérez-Gussinyé & Reston 2001), which would then provide pathways for water to enter the mantle and enable the serpentinization of the peridotite. In analogy to the previous discussion on the absence of a HVLC at that margin, reduced mantle velocities may not have been resolved with the set-up of the experiment.

In the centre of the JMMC, crustal velocities vary between 6.2 and 6.8 km s^{-1} (Mjelde *et al.* 2008). Velocities of 5.8–6.2 km s^{-1} that are observed in the upper crust of Greenland (Hermann & Jokat 2016) are not present. This may indicate that the upper crust may have been completely removed during the rifting. However, some lower velocities of 6.0 km s^{-1} are reported on line 4 of the JMKR-95 survey (Kodaira *et al.* 1998*b*) a little further to the north on the JMMC (Fig. 9a). These velocities would be more compatible with the upper-crustal velocities of central East Greenland.

The transect compiled for the eastern JMMC and Møre margins (Fig. 10b) is conjugate within 75 km, and the reconstruction is for 49 Ma corresponding to Chron C22. The oceanic crust at that age has a similar thickness (5 km) and velocity structure on either side. However, some asymmetry in the spreading rate becomes visible with a wider zone of oceanic crust on the Møre side. In addition, the oceanic crust on the Møre side becomes thicker towards the COB when compared to the JMMC side. The initial oceanic crust has a thickness of 8 and 11 km at the eastern JMMC and Møre margins, respectively. These thickness differences can probably be attributed to deviations from the exact conjugacy. Landward of the COB, a high-velocity lower-crustal body is observed on either side. Beneath the Møre marginal high, the HVLC is 4 km thick: however, the velocity model for the JMMC does not show a sharp boundary between regular lower crust and a HVLC, but, rather, combines them into a single layer with a maximum thickness of 7 km. The extended continental crust beneath the Møre Basin is modelled as a single layer, the thickness of which is less than 7 km in a 240 km-wide zone landward of the COB. Crustal velocities range from 6.1 to 6.7 km s^{-1}, which is compatible with the JMMC, where the maximum crustal thickness is 7 km outside the zone with the HVLC. While the crustal extension at this margin pair is distributed over a wide zone, the break-up occurred close to the western edge of this zone. The HVLC is observed beneath the entire length of the hyper-extended crust, but disappears beneath the western part of the JMMC that was probably much thicker at the time of the Early Eocene break-up between the JMMC and Norway. This thicker crust could then have prevented magma spreading westwards.

The first conjugate transect to the north of the Jan Mayen Fracture Zone is composed of lines AWI 20030400 (Voss & Jokat 2007) off East Greenland, and a combination of line 10 of the OBS 2003 survey (Breivik *et al.* 2009) and line 5 of the Vøring 96 experiment (Mjelde *et al.* 1998) at the Vøring margin (Fig. 11c). The lines are reconstructed for Chron C22 (49 Ma) and they are conjugate within 30 km. Unfortunately, the two Norwegian profiles cover only the outer part of the margin, and do not extend across the Trøndelag Platform and further to the coast. At Chron C22, there is a good match of the velocities and thickness (10 km) of the oceanic crust on either side of the transect.

Problems arise when the COB is looked at in more detail (Fig. 11c). At the Vøring margin, the initial oceanic crustal thickness is 18 km, compared to 24 km on the Greenland side when the COB of Funck *et al.* (2014) is used. Employing the original COB interpretation of Voss & Jokat (2007) on the Greenland side would make for an extreme asymmetry of oceanic crustal accretion, with a 30 km-wide zone of oceanic crust accreted prior to C22 on the Greenland side compared to more than 100 km off Norway. The oceanic crustal thickness at the original COB of Voss & Jokat (2007) is 14 km, which does not fit with the conjugate

Fig. 11. (**a**) Conjugate transect between NE Greenland and the Lofoten margin constructed from results of gravity inversion (Haase *et al.*, this volume, in press) and the sediment thickness compilation of Hopper *et al.* (this volume, in prep). P-wave velocities are not known along the transect. (**b**) & (**c**) P-wave velocity models along conjugate transects between East Greenland and the Vøring margin. Line AWI 20030300 (after Voss *et al.* 2009), lines 3 (after Mjelde *et al.* 1992) and 1 (after Mjelde *et al.* 1993) of the Lofoten 88 survey, line AWI 20030400 (Voss & Jokat 2007), line 10 of the OBS 2003 survey (after Breivik *et al.* 2009), and Vøring 96 line 5 (after Mjelde *et al.* 1998). The inset map displays the age of the oceanic crust (after Gaina 2014; cf. Gaina *et al.*, this volume, in review), together with the line locations (blue lines) and flow lines (dashed white lines). Thin solid lines mark selected isochrons. Abbreviations: COB, continent–ocean boundary; HVLC, high-velocity lower crust; JMMC, Jan Mayen microcontinent; LR, Lofoten Ridge; orig., original interpretation; SSB Skomvær Sub-basin.

margin either unless there was magmatic addition after the accretion of the crust, which could account for the difference. The mismatch between line AWI 20030400 and the conjugate Vøring margin was also discussed by Voss & Jokat (2007). They argued that anomaly C22 is the oldest true seafloor spreading anomaly along their Greenland line. Magnetic anomalies landward of C22 are interpreted to relate to intrusions into stretched continental crust. Furthermore, Voss & Jokat (2007) argued that their magnetic data show that anomalies C24A–C21 terminate against the East Greenland margin. This would be compatible with a north–south propagation of rifting between Shannon Island and the Jan Mayen Fracture Zone. If this model holds, the COB on the Norwegian margin would need to be moved further seaward and the oldest identified magnetic anomalies be questioned. Given the wealth of data on the Vøring margin, it is easier to attribute problems in plate reconstructions to a lack of sufficient data at the conjugate Greenland margin. However, it is worthwhile to mention that magnetic anomalies C24B–C23 become progressively more diffuse to the south at the Vøring margin (Olesen *et al.* 2010), not too dissimilar to what is observed off Greenland. Based on the velocity models alone, it is difficult to pinpoint the exact location of the first true oceanic crust on either of the lines, which is why the real question might be whether or not the diffuse magnetic anomalies could be highly stretched and intruded continental crust or associated extrusive volcanic rocks that may overlie such crust.

Both lines display a HVLC (Fig. 11c), although it is much thicker on the Greenland side (up to 16 km) than beneath the Vøring margin (up to 9 km but mostly *c.* 6 km). Most of the Mesozoic crustal extension is taken up by the Vøring margin. At the eastern end of the transect, some 160 km landward of the COB, the continental crust excluding the HVLC is only 8 km thick, while the Greenland crust has a thickness of 25 km at a similar distance.

Moving northwards, line AWI 20030300 (Voss *et al.* 2009) is conjugate within 60 km of lines 1 (Mjelde *et al.* 1993) and 3 (Mjelde *et al.* 1992) of the Lofoten 88 survey (Fig. 11b). While the two Norwegian lines are located northwards of the Bivrost Lineament and, hence, are part of the Lofoten margin, line AWI 20030300 would reconstruct to a position just to the south of the lineament. Hence, along-margin variations across the lineament may affect the conclusions drawn from this conjugate transect. At 53 Ma, the oceanic crust has a thickness of 12 km on either profile. While the Moho deepens landward on the Greenland side, its depth does not change over a distance of 90 km at the Lofoten margin. The original interpretation of the COB on line AWI 20030300 by Voss *et al.* (2009) seems to be in better agreement with the conjugate Lofoten margin than the revised COB by Funck *et al.* (2014). Since the lower-crustal velocity increases steadily from west to east between km −90 and km −10 (Fig. 11b), this zone fits better the characteristics of a COT than those of oceanic crust. Some HVLC is observed in this transition zone, while it is less clear whether there is such a high-velocity lower-crustal body on the Lofoten lines. Instead of a gradual velocity increase, there is a sharp velocity change some 15 km landward of the COB. The Moho deepens gradually on the Greenland side. In contrast, some local shallowing of the Moho is indicated beneath the Lofoten Ridge. The ridge separates the Skomvær Subbasin from the Vestfjorden Basin with a maximum sediment thickness of 7 km. The depth of the sedimentary basins on the Greenland margin is poorly resolved, as explained earlier. They are possibly deeper than the 5 km indicated in Figure 11b and might be as deep as 14 km, as suggested in the original interpretation of the velocity model (Voss *et al.* 2009).

Given the lack of seismic refraction data covering the deep and wide basins beneath the NE Greenland Shelf, alternative data have to be used to illustrate the conjugate Lofoten–NE Greenland margins in their full width. Figure 11a shows such a transect, which is based on three datasets: the sediment thickness map of Hopper *et al.* (this volume, in prep); the Moho depth obtained from gravity inversion (Haase *et al.*, this volume, in press); and the COB of Funck *et al.* (2014). While this transect does not allow verification of crustal features based on their seismic velocities, some general remarks on the crustal architecture are possible. The conjugates are reconstructed for 47.3 Ma (C21), where the oceanic crust has a thickness of 6 km. At that position, the basement is slightly deeper on the Lofoten profile, which is related to a thicker sediment load. The oceanic crust thickens towards the COB, where it has a thickness of 8 and 10 km at the Greenland and Lofoten margins, respectively. This thickening indicates a larger magma supply following break-up and would be compatible with some HVLC in the adjacent COT. This is also consistent with the HVLC observed on line AWI 20030200 (Fig. 6b) that lies within 5 km of the transect.

The NE Greenland margin is characterized by a 410 km-wide zone with extended continental crust (Fig. 11a). At the Lofoten margin, the continental crust thickens from 10 to 30 km over a distance of just 100 km. In contrast to the Vøring margin further to the south, the polarity of the margin has changed from a wide margin on the Norwegian side (Fig. 8c) and a narrow margin off East Greenland (Fig. 6d) to the opposite configuration (Fig. 11a). Beneath the NE Greenland Shelf, two deep basins are encountered: the 7 km-deep Thetis Basin and the 17 km-deep North Danmarkshavn Basin (Fig. 11a). While the crust beneath the Thetis Basin still has a thickness of 11 km, it thins to 6 km beneath the North Danmarkshavn Basin and has to be considered as hyper-extended. Unfortunately, there is no information on the crustal and mantle velocities beneath the basin that could provide information on the existence of a HVLC or partially serpentinized mantle. This could provide additional constraints on the nature of the HVLC at the Vøring margin, where its composition is controversial (cf. Mjelde *et al.* 2009).

Conclusions

The review of the seismic refraction data has shown that there is a wealth of profiles that provide information on the continental margins bordering the NE Atlantic Ocean. The margins are, in large parts, characterized by excess magmatism associated with the volcanic break-up in the Early Eocene. This magmatism is evidenced by SDRs on seismic reflection sections, while refraction data image mafic intrusions that are accompanied by an increase in crustal velocities. A HVLC is also often observed, which is commonly explained by magmatic underplating or as a result of sill intrusions into the lower crust (White *et al.* 2008) at the distal parts of the margins. However, high velocities are also observed below more proximal parts of the margins, such as beneath the southern Rockall Basin and the Porcupine Basin, where they are interpreted as partially serpentinized mantle (Shannon *et al.* 1999; O'Reilly *et al.* 2006). At the mid-Norwegian margin, interpretations vary from magmatic underplating to eclogite bodies and partially serpentinized mantle (cf. Mjelde *et al.* 2009; Lundin & Doré 2011). Crustal thinning down to a few kilometres (e.g. Fig. 8b) is observed beneath the rift basins that are generally underlain by material with seismic velocities in excess of 7.0 km s^{-1}. As stated above, the lithological interpretation of these velocities is often controversial.

The Iceland plume with its associated excess magmatism has shaped the continental margins bordering the NE Atlantic. Along the Greenland–Iceland–Faroe Ridge (Fig. 5), the crustal thickness is generally greater than 30 km (Richardson *et al.* 1998; Holbrook *et al.* 2001). The thickness of the initial oceanic crust formed at break-up decreases southwards. However, even some 1200 km away from the axis of the Greenland–Iceland Ridge, the igneous crust at the SE Greenland margin is still 16 km thick (Holbrook *et al.* 2001) (Fig. 2c). Magmatism to the north of Iceland was more restrictive when compared to the south. At the central Møre margin, oceanic crust at the time of break-up is only 11 km thick and thins to 5 km towards the extinct Aegir Ridge (Fig. 8c) (Breivik *et al.* 2006). Later, when the spreading moved from the Aegir Ridge to the Kolbeinsey Ridge, the thickness of the oceanic crust increased again to 9 km to remain fairly constant around this value (Figs 6e & 9b) (Kodaira *et al.* 1998*a*; Hermann & Jokat 2016). A significant increase in the amount of magmatic addition is noticed north of the Jan Mayen Fracture Zone with an initial oceanic crustal thickness of 18 km at the Vøring margin (Fig. 8b) (Breivik *et al.* 2008) or even up to 25 km on the conjugate NE Greenland margin (Fig. 6d) (Voss & Jokat 2007). Similar to the area south of Iceland, the amount of magmatism decreases with distance to the plume. On the Greenland side, the northernmost evidence of a HVLC is found just to the south of the East Greenland Ridge (Fig. 6b) (Voss *et al.* 2009). On the conjugate Norwegian side, there is no evidence of a HVLC at the Lofoten margin, even though the region was still affected by extrusive magmatism, including SDRs (Fig. 1).

The overall architecture of the rifted margins in the NE Atlantic Ocean reveals a pronounced asymmetry of the conjugate margin segments. South of the Bivrost Lineament, the wide margins with deep rift basins are located on the European side, while the conjugate East Greenland margin is comparatively narrow. This pattern changes north of the Bivrost Lineament, where the Lofoten margin is narrow, while the conjugate NE Greenland Shelf with the deep Danmarkshavn and Thetis basins widens northwards.

This review has shown that there are still many questions that the presently available data cannot answer. In particular, there is a shortage of truly conjugate transects that can help to fully understand the complex rifting history in this part of the Atlantic. There is a lack of continuous profiles extending over the entire width of, in particular, the wide margins (the Hatton–Rockall area, and the Vøring and NE Greenland margins) from the proximal to the distal zones into truly oceanic crust. The mid-Norwegian margin is studied by a dense net of seismic refraction lines that are generally short. Even if they can be merged into a continuous profile, there are often substantial deviations at the line intersections (e.g. Figs 8b & 11c). In NE Greenland, seismic data acquisition is impeded by the presence of sea ice, and north of the East Greenland Ridge there are no refraction lines to image the margin there. The mid-Norwegian margin is generally considered as well studied, while the coverage on the conjugate East Greenland margin is less dense. However, some modern refraction lines are available for East Greenland and cover that margin in its full width (Voss & Jokat 2007; Voss *et al.* 2009; Hermann & Jokat 2016). In contrast, the most recent lines published on the Lofoten margin were acquired in 1988 (Mjelde *et al.* 1992, 1993) with a poor resolution in the COT due to a wide receiver spacing. Velocity models there indicate a sharp transition from oceanic to continental crust, in contrast to wider COTs observed at magma-rich margins studied with more receivers, such as the Hatton margin (White & Smith 2009) or the NE Greenland margin (Voss & Jokat 2007, 2009; Voss *et al.* 2009). As reflection data and potential field data alone are not necessarily sufficient to define the COB, some modern OBS experiments would be desirable to confirm or refine the COB at the Lofoten margin and also at the Vøring margin, at the transition from the Vøring to the Møre margin, and probably many more places.

Similarly, at the SE Greenland margin, there remains uncertainty about the location of the COB, as White & Smith (2009) questioned the interpretation of some weak magnetic anomalies as true seafloor spreading anomalies (Larsen *et al.* 1994; Larsen & Saunders 1998). Using the conceptual model of Quirk *et al.* (2014) for the break-up of volcanic margins, some of the previously interpreted oceanic crust at the SE Greenland margin may, indeed, be thinned and intruded continental crust that is overlain by a broad zone of subaerial SDRs. This may create a velocity structure that looks more oceanic than continental, and would reduce the asymmetry in the initial accretion of oceanic crust. However, a verification of the model in Quirk *et al.* (2014) would require seismic reflection data imaging down to the Moho, preferentially on both the SE Greenland and the conjugate Hatton margins. Poorly defined COBs may be the reason for the apparent asymmetry in the production of early oceanic crust in many parts of the NE Atlantic.

The margins of the JMMC are only poorly sampled so far. Some refraction profiles are available for the northern JMMC but leave some questions open, in particular about the western margin. A recently published line at the East Greenland margin (Hermann & Jokat 2016) conjugate to the JMMC shows some indication for the presence of a HVLC with no equivalent at the JMMC (Kodaira *et al.* 1998*a*). Since the two lines are not completely conjugate, the differences could relate to along-strike variations at the margin. Alternatively, there could be problems in the velocity models. The southern limit of the JMMC is not defined with any confidence and there is no seismic refraction data available to provide information on this highly segmented part of the microcontinent.

Another major data gap is the sheared NE Greenland margin to the north of the East Greenland Ridge. This area is ice-covered year-round in contrast to the much better studied conjugate SW Barents Sea margin. Of particular interest here would be a conjugate to the Vestbakken Volcanic Province that is characterized by volcanic rocks and a HVLC (Fig. 7c) (Libak *et al.* 2012). Conjugate to this province, seismic reflection data show some strong, but discontinuous, reflectors that might be SDRs (cf. Geissler *et al.*, this volume, in review). Refraction data could provide information on the deeper crustal velocity structure and verify whether or not the margin is similar to the Vestbakken Volcanic Province.

Another issue with many seismic refraction lines crossing the deep sedimentary basins off NE Greenland, mid-Norway and in the SW Barents Sea is the clear identification of the basement, as velocities of Mesozoic and Palaeozoic sedimentary rocks can be similar to crystalline basement when deeply buried. This can leave some uncertainty with respect to crustal thinning factors, which are important for basin modelling and deformable plate reconstructions. Some of the ambiguity can be reduced by having coincident deep seismic reflection data of high quality available for the modelling of the refraction data. This may require some collaboration between industry and academia, as there are excellent industry seismic reflection data acquired along the margins that would benefit from adding refraction data. A better integration of potential field and seismic reflection data could help to resolve issues related to sparse seismic refraction data and difficulties in imaging deep structures.

While magma-poor margins are fairly well understood by now, thanks to extensive studies including drilling at the conjugate Iberia–Newfoundland margin pair (cf. Peron-Pinvidic *et al.* 2013), there has been less focus on the rifting processes at magma-rich margins (Quirk *et al.* 2014). The NE Atlantic with the North Atlantic Igneous Province is a natural laboratory to study magma-rich margins. However, despite decades of seismic and geophysical data acquisition, there is still a lot of uncertainty on first-order structures, such as the location of the COB. This is why some key conjugate refraction transects should be acquired in the NE Atlantic. These should be truly conjugate, cover the entire width of the margins, have good coincident reflection data available, use a similar instrumentation and experimental set-up on either side, and would be analysed in a consistent way with the same techniques, ideally by the same person.

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